Dynamical Climatology

A Budyko-Sellers model for study of ocean-atmosphere interaction.

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DCTN 26

July 1985

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DYNAMICAL CLIMATOLOGY TECHNICAL NOTE NO. 26

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1. Introduction

Energy balance climate modelling takes a very simple approach to climate modelling, nothing more than the conservation of energy - the temperature distribution depends simply on the way energy comes in and leaves the system and the way it is moved around. Zero-dimensional global or hemispheric energy balance models (EBMs) can give a simple look at phenomena such as the greenhouse effect; one-dimensional EBMs (zonally-meaned or "radiative-convective" with the vertical dimension only) can be used to give a clearer understanding of the processes going on in the real world or in general circulation models (GCMs). They are necessarily based on highly simplified parametrizations of the real physical processes, and any results from them should be treated with care: they can be simultaneously unreasonably sensitive or unreasonably insensitive to these parametrizations. Their merit, apart from computational cheapness, is simplicity - it should always be plain precisely what is happening in an EBM, and what its approximations and inaccuracies are, in a way not true for a GCM. It is also possible to explore their possibilities more exhaustively by varying almost every aspect, although tuning to the present climate provides strong constraints - not necessarily the right ones, as tuning an EBM does not mean it will give correct reactions to altered external forcing - which limit their use in studies of climate change.

However, as models, and cheap and simple models, they may be useful in looking at aspects of the behaviour of more complex ones, and this one has been devised with the testing of schemes for accelerated convergence of coupled ocean/atmosphere GCMs in mind. It is based very largely on the discussion of Budyko-Sellers models, i.e. zonal-mean EBMs, by North, Cahalan & Coakley (1981), which is recommended for reading on more general aspects of energy balance modelling, and, for the coupled aspects, on a paper by Dickinson (1981) and previous work in the Dynamical Climatology Branch by D.L. Roberts on a version of Dickinson's numerical model, which is globally-meaned with 3 layers in the vertical.

Obviously at least 2 layers in the vertical are needed to look at ocean/atmosphere coupling. However, the first version of the model had one layer only, representing the combined effects of the atmosphere and oceanic mixed layer (OML), which simplifies writing and testing the numerical scheme, and makes it easier to follow the model's idealizations. A 2-layer version resolves the atmosphere and OML, and 3-layer versions allow also for what are known as the ocean intermediate waters, in two slightly different ways. In line with the conceptual and numerical simplicity desired, the model is zonally meaned and its parametrizations are chosen for simplicity and convenience.

Section 2 of this note describes the model's physical and numerical aspects - the general reader may skip the latter - Section 3 then describes the results from the various versions and Section 4 contains a brief summary and conclusions.

2. Description of the model

As an energy-balance model, the basic equation must be an expression of the 1st law of thermodynamics. For each layer we take the diffusion equation for a spherical shell (see, for example, North et al):

$$C \frac{dT}{dt} + \frac{d}{dx}\left\{(1 - x^2)\, D(x,t) \frac{dT}{dx}\right\} = Q(x,t) \qquad (2.1)$$

where x is the sine of latitude, the basic space variable, C the layer's heat capacity per unit area, D an effective diffusivity, though its dimensions are not L^2T^{-1} (the factor $(1-x^2)$, allowing for the contraction of latitude circles towards the poles, could be absorbed into this) and Q the heat gained or lost by radiation to space or exchange with other layers.

The assumptions come in the treatment given to C, D and Q, describing just how the layer stores, transports and gains or loses heat. The simplest treatment of C, which I have adopted, is to keep it a constant for each layer. Thompson & Schneider (1979) advocate the use of a latitude-dependent and seasonally-varying heat capacity for the oceanic mixed layer, taken from climatological data, so as to simulate more accurately the seasonal cycle. Given the crudity of the treatment of the ocean, and of the air-sea fluxes, the specification of a large amount of data from climatology and the calculation or specification of a thermocline temperature for the seasons when the OML is deepening seemed an unnecessary complication, as well as tying the model too much to present-day climate. Nor have I incorporated a time-constant latitude dependence, though this would be a fairly simple modification. The values used are, in W yr $m^{-2}K^{-1}$, 0.45 for the atmosphere, 10 for the OML or the 1-layer version of the model and 100 for the intermediate waters, equivalent to 1.5×10^7, 3.3×10^8 and 3.3×10^9 $Jm^{-2}K^{-1}$, or about 3.5, 80 and 800 m of water respectively. These values are taken from Dickinson (1981) and were also used in Roberts's numerical work. The atmospheric value has been multiplied by a factor 1.4 to allow for the oceans' only covering 0.7 of the earth's surface, so that the values "per unit area" are in fact "per unit area of ocean". The program has been written to make it easy to change these values. No attempt has been made to allow for the zonal variation of the water-covered fraction of the earth's surface, or to simulate any "temperature over land".

For the diffusivity, the simplest possibility, representing it as a constant, has again been used. For the 1-layer model, the value of 0.649 $Wm^{-2}K^{-1}$, taken from North et al was used, while for the 2-layer the sum of the diffusivities for atmosphere and OML comes to this, with the proportion in each layer being set in a subroutine call. (Because D in (2.1) has been multiplied by C compared with a "true" diffusivity, the total heat transport depends only on the total D and the temperature gradients. To convert to a "true" diffusivity with dimensions L^2T^{-1} multiply by the square of the earth's radius and divide by the heat capacity of the layer concerned, so that 0.649 $Wm^{-2}K^{-1}$ in the OML is equivalent to 8×10^4 m^2s^{-1}.) For the 3-layer model values were found by tuning, as described below. A constant value corresponds to purely random "diffusive" transport, which is certainly not the case for either the atmosphere, where baroclinic instability means that the diffusivity

increases with temperature gradient, or the ocean, with its highly organized circulations. Stone (1973) put forward and verified an analytically-based expression proportional, amongst other things, to the hemispheric mean meridional temperature gradient divided by the Coriolis parameter, i.e. flux proportional to the square of temperature gradient. Gal-Chen & Schneider (1976) compared the effects of various flux parametrizations, and found such sophistications as making the diffusivity increase with $q(T)/T^2$, where q is the saturated humidity mixing ratio, (to mimic the effects of the Clausius-Clapeyron relation on latent heat transport), or including prescribed advection, tended to give the system a less natural behaviour. However they did feel that a square-gradient law (in their case with constants derived from observations) gave a more realistic model, particularly with regard to changing external forcing. For the sake of simplicity of coding and keeping the system linear and analytically tractable, I have not followed this, as it would have little effect on the convergence of the system and in any case would only be appropriate for the atmospheric fluxes: parametrizations of the oceanic fluxes have hardly been discussed in the literature, being less well known in reality and more complex in that they are dependent on external dynamical forcing. In fact, the meridional fluxes of an EBM are extremely insensitive to transport parametrizations - Stone (1978) has shown that as long as the distribution of incoming radiation is correct an EBM ignoring meridional variation of albedo and temperature gets them right within the limits of observational error. So we may tune using only the temperature distribution, and in general this is not difficult (though special measures would have to be taken to get the distribution accurate in the tropics - see Lindzen and Farrell, 1977). The model has been written so that introducing a variable diffusivity, of any form, would not require many alterations.

The term Q(x,t) consists of solar radiation absorbed, long-wave radiation lost, and energy transferred between layers. Following North et al, the solar constant is taken as 1340 Wm^{-2} and the proportional distribution of solar radiation and the earth albedo are approximated by 2nd order Legendre polynomials in sin(latitude) taken from symmetrized Northern Hemisphere data, i.e. Northern hemisphere data are used for both hemispheres, with a half-cycle phase difference. Both annual mean and seasonally varying forms have been used. It would, however, be simple to use a more precise function, or to introduce, say, albedo/temperature feedback. Figs 1 and 2 compare the correct values, taken respectively from a diagram attributed to Gruber (1978) in J. Slingo (1982), and from A. Slingo (1982), with the harmonic approximations.

Outgoing radiation uses a linearized form, A + BT, with A = 203.3 Wm^{-2} and B = 2.09 $Wm^{-2}K^{-1}$, again taken from North et al. Some authors (e.g. Gal-Chen & Schneider, 1976) have used much more complex forms, but given the level of approximation inherent in defining it in terms only of a zonal mean temperature, this hardly seems worthwhile: obviously any properly tuned formula will be to first order equal to the same linear formula. In fact Manabe & Wetherald (1967) have shown that a linear dependence may be surprisingly accurate, as the more-than-linear dependence given by the Stefan-Boltzmann law (and also by the decrease in moist adiabatic lapse rate with temperature) is counteracted by the water vapour greenhouse effect. The proportions in which both incoming and outgoing radiation affect the atmosphere and OML respectively are set by the user of the model. (There are two ways of obtaining a value for B: tuning so that A + BT best describes the variation in outgoing infrared as local temperature varies over the globe for present climatic conditions; or so that it describes the dependence of total

planetary emission on planetary mean temperature during a climate change due to, e.g., change in the solar constant. The original work of both Budyko and Sellers took a value found in the former way and used it for the latter process, a method which estimates climate sensitivity simply but unreliably -North et al point out that it would be doubled by a 2% change in global mean fractional cloud cover for every 1 K rise in global mean temperature. The value used here, from North et al, was produced by the former process. Dickinson (1981) and Roberts however both used a value 2.4 Wm^{-2} K^{-1}. This was obtained by multiplying by 1.4 the midpoint of the range 0.9-2.5 Wm^{-2} K^{-1}, estimated as the likely range for the latter process from AGCM climate sensitivity studies. Therefore our results are not completely comparable, though convergence rates at a point are similar, the slowing of convergence in this model due to a smaller value of B being counteracted by the additional effect of diffusion in the added horizontal dimension.)

Heat exchange between layers, when there is more than one, has been linearized also, i.e. $\lambda_{12}(T_1-T_2)$ with exchange coefficients 45 $Wm^{-2}K^{-1}$ between OML and atmosphere and 1 $Wm^{-2}K^{-1}$ between OML and intermediate waters. The former value is taken from Dickinson (1981) who obtained it from sensitivity experiments with the NCAR GCM, which also suggested that the linearization assumption is accurate to within around 7%; it was also used in Roberts's work. It is possibly too high (Bretherton, 1982, suggests an average value of about 30 $Wm^{-2}K^{-1}$ - note that his comments on heat losses from large scale SST anomalies refer to linearized forcing back to the equilibrium SST, not to the air temperature, and the effects mentioned are in fact allowed for in this model) but as it is rather faster than any other processes involved, the system is not sensitive to its precise value. The linearization assumption is less appropriate between OML and intermediate waters, where most heat interchange occurs through specialized processes such as high latitude winter mixing and equatorial upwelling; in fact there is no need of such a term in the version of the 3-level model which has a prescribed circulation, while the version which is purely diffusive has a value derived by tuning as described below. For a model with meridional variation, there is also the possibility of making the coefficients latitude dependent - they are larger in reality at higher latitudes due to lower stability and higher wind-mixing, and in the case of OML/intermediate water coupling, to the occurrence of convective overturning, and then less again around the pole, thanks to the insulating effect of ice and the stability provided by the Arctic halocline. For simplicity, this has not been done (other than the provision of convective overturning in the 3-layer models, the only natural way of ensuring that the lower oceanic layer is colder).

It should be noted that the "air temperature" is taken to be a surface temperature for calculation of air-sea fluxes and radiative cooling rates although it is calculated using a heat capacity appropriate to the whole atmosphere. It can be thought of as related to a vertically meaned potential temperature. Some authors use a nominal 500 mb temperature as their working variable and add on a constant, representing a prescribed lapse rate to the surface, whenever actually using it to calculate an energy exchange. This gives equivalent results to the simpler formulation used here.

Use of a constant diffusivity makes the Legendre polynomials eigenfunctions of the system, and North et al did their analysis entirely in terms of them. However, coding the model spectrally would make it difficult to run with variable diffusivity or with a circulation, and would hamper the

use of something like ice/albedo feedback, which is by its nature local. A grid-point scheme also has the advantage that the harmonics can be used to check the results: but not vice versa. The finite difference scheme used is the Crank-Nicholson (see Richtmyer and Morton, 1967, pp 185-201) which assumes

$$\frac{dT}{dt} \approx \frac{T_i^{n+1} - T_i^n}{\Delta t} \qquad \frac{d^2T}{dx^2} \approx \frac{T_{i+1}^{n+1} + T_{i-1}^{n+1} - 2T_i^{n+1} + T_{i+1}^n + T_{i-1}^n - 2T_i^n}{2\Delta x^2} \qquad (2.2)$$

This is one of the standard methods for the diffusion equation, and is stable, robust and the most accurate (second order in both space and time) of the reasonably simple schemes. Although it is always a stable treatment for the pure diffusion equation, there is also a negative feedback in the BT term of the longwave radiation and the λT terms of the exchange between layers, and this will give instability for too long a timestep because throughout the timestep it is effectively fixing the heat loss at a value appropriate at the start, and if it goes on too long it will overshoot the equilibrium value and end up even further away. This begins when the timestep exceeds two e-folding times of the layer, the smallest value of which is for the atmosphere with C_a = 0.45 W yrs $m^{-2}K^{-1}$ and λ_{am} = 45 $Wm^{-2}K^{-1}$: $^1/_{50}$ yr, or about a week. To get an accurate description of the time evolution of a layer the timestep should be rather shorter than required for stability.

A "box" scheme is used, with each value being thought of as representative of values over a latitude belt: the physical boundary condition of zero heat flux across a pole then takes the mathematical form of zero diffusivity on the polar side of an extremal belt, which is intuitively satisfying and easy to code.

As the scheme is semi-implicit, it does not give the temperatures for the next timestep directly. Substituting (2.2) into (2.1) gives:

$$- D_{j+1} T_{j+1}^{n+1} + Q_j T_j^{n+1} - D_j T_{j-1}^{n+1} = - D_{j+1} T_{j+1}^n + F_j T_j^n - D_j T_{j-1}^n + \frac{\Delta t}{C} H_j \equiv E_j$$

with D_j representing the effective diffusivity between points j and j-1, Q_j and F_j taking into account effects of diffusivity and temperature-dependent radiation, and H_j the prescribed part of the local radiative gain or loss, and any transfer between layers.

This is a tridiagonal matrix equation, thanks to the diffusion terms which make the temperature increment at each point depend on the temperatures at adjacent points. The matrix to be inverted is symmetric and for some parameter values close to the identity, and one numerical scheme was tested which takes advantage of this. However, the inversion normally used is a general one for all tridiagonal systems:

given
$$\begin{pmatrix} b_1 & d_1 & 0 & & & 0 \\ a_2 & b_2 & d_2 & & & 0 \\ 0 & & & & & \\ & & & a_{m-1} & b_{m-1} & d_{m-1} \\ 0 & 0 & & 0 & a_m & b_m \end{pmatrix} \begin{pmatrix} T_1 \\ T_2 \\ \\ T_{m-1} \\ T_m \end{pmatrix} = \begin{pmatrix} e_1 \\ e_2 \\ \\ e_{m-1} \\ e_m \end{pmatrix}$$

one can solve in 3 steps: first find w:

$$w_1 = \frac{d_1}{b_1}, \quad w_i = \frac{d_i}{b_i - a_i w_{i-1}} \qquad i=2,\ldots,m-1$$

then g:

$$g_1 = \frac{e_1}{b_1}, \quad g_i = \frac{e_i - a_i g_{i-1}}{b_i - a_i w_{i-1}} \qquad i=2,\ldots,m$$

and finally T: $T_m = g_m$, $T_i = g_i - w_i T_{i+1}$, $i=m-1,\ldots,1$

Since the w_i depend only on the elements of the matrix, and are constant in the case of constant diffusivity, they are calculated at the start of the relevant subroutine, and then at each timestep the e_i are found from the radiational balance and current temperatures, and then g_i and then T_i. As mentioned above, to alter the model to variable diffusivity would be simple enough though it would increase the number of computations necessary.

3. RESULTS

a. 1-layer version

For the 1-layer model a simple analytical solution exists for the equilibrium with constant diffusivity and 2nd order harmonic forcing, which is also of 2nd order harmonic form. Fig 3 shows numerical results for annual mean forcing at various spatial resolutions (since the forcing is symmetric, so is the equilibrium): it can be seen that agreement is good even with only 10 points in each hemisphere. (Note that in all diagrams, unless indicated otherwise, the abscissa is sin(latitude).)

Fig 4 shows the effect of altering the basic tunable constants, in each case by ±10% from the standard value, on the equilibrium with annual mean forcing. (Altering the heat capacity has no effect on the final equilibrium, as it only enters as a scaling factor for time, though this would of course alter the equilibrium seasonal cycle for the seasonally forced case.) The response is generally as one would expect intuitively or mathematically, although for alterations of the solar constant not only does the mean temperature increase or decrease with the solar constant, but so does the equator-pole difference (and higher harmonics) which is the reverse of the behaviour observed in GCMs and believed to occur in reality. This is because

in the EBM we simply alter the incoming radiation everywhere in proportion to its previous value, and so most at the equator, while in reality this effect is counteracted by albedo-temperature feedback (which,though it changes temperatures globally, does so most nearest its source area) and by change in strength of the polar inversion, which increases the temperature change observed at the surface. Indeed Gal-Chen and Schneider (1976) used consistency of increase of pole-to-equator temperature differences with solar constant in the absence of albedo-temperature feedback as a criterion for judging models. It should be made clear that the model is not intended to give a "realistic" response to perturbations, as this would require allowing accurately, implicitly or explicitly, for every process significant in climate change. A model merely tuned to the current climate cannot be expected to do this, and using detailed parametrizations derived from a more complex model should simply reproduce the behaviour of the more complex model (which may be useful if one wants to check what the dominant factors are: e.g. Held et al, 1981).

The simplified seasonal forcing (i.e. albedo and insolation) used produces a seasonal response of the same harmonic form, sinusoidal in time and proportional to distance from the equator (Figs 5 and 6), which is not of course very realistic in the tropics. It also gives an overall warming, due to the positive seasonal correlation between albedo and insolation - at high latitudes, when there is more sun, in summer, there is less snow and ice and a higher proportion is absorbed. This is a "real" effect, and indeed the variation of its strength with variations in the earth's orbit (magnified by albedo/temperature feedback) is the basis of the Milankovitch theory of the ice ages. Fig 6 shows that the seasonal variation of temperature is reproduced generally to within 0.1 K of the analytic solution, even at the fairly coarse resolution used for that run. Results from sensitivity experiments in which the resolution was increased (not shown) implied that truncation error in space and in time are each responsible for about half the magnitude of the error, giving respectively an error in the mean and in the seasonal cycle. Rounding error was also significant, since for those particular parameter values the inversion problem is ill-conditioned, and the approximate alternative numerical scheme referred to above gave significantly better answers. The other main residual effects of the seasonal cycle are a small reduction in the overall equator to pole gradient (the albedo/insolation feedback originating in high latitudes) and a small semiannual cycle in the planetary mean temperature, shown in Fig 7: as can be seen, the seasonal variation is well shown, though there is an error in the time mean of the same order (which again was reduced by increased resolution in space.) It is not of course suggested that this semiannual cycle has any physical meaning, but the model's reproduction of all the analytic modes in this simple case inspires confidence in its performance in more complex cases.

In spite of the small "errors" that are introduced into the results in Figs 4-7 by the use of only 10 space steps to a hemisphere, this value was also used for all integrations mentioned below (the ill-conditioning mentioned above does not then occur when the timestep is reduced to a value suitable for integration of the models which resolve the atmosphere.)

The transient behaviour of the analytic model is that the various harmonics decay independently, the higher the faster, so the overall decay rate at a point at a given time would depend on the harmonics most out of equilibrium there. Fig 8 shows the convergence of the numerical solution at a

near-equatorial point in a run with seasonal forcing and initial conditions isothermal, near to the equilibrium mean temperature, and also the analytic convergence rate for the first 3 harmonics, drawn through the position for the end of the first year. With the 0^{th} harmonic (planetary mean temperature) and the 1^{st} harmonic (the seasonal cycle, sinusoidal in time and proportional to distance from the equator) near equilibrium, we would expect convergence to be dominated at first by the 2^{nd} harmonic (parabolic in form and without seasonal variation , representing equator-to-pole mean temperature variations), and then, as this converged, the rate to be more influenced by the 1^{st} and 0^{th} harmonics. More detailed analysis shows that the 0^{th} harmonic replaces the 2^{nd} as the largest during the 6^{th} year, the 1^{st} never being as significant. This seems to agree quite well with the actual numerical behaviour, and suggests that the tailing off of convergence is not just the effect of rounding error becoming more important as the changes approach the limit of accuracy of the machine. (The "equilibrium" values quoted above were all run for at least 8 e-folding times of even the 0^{th} harmonic, and in most cases printing out intermediate values confirmed convergence had occurred.)

b. 2-layer version

Extending the model to 2 layers, as described above, requires the specification of heat transfer, of heat capacities for the two layers, and of the proportion of incoming radiation, outgoing radiation and diffusivity in each. Development was carried out with the latter processes affecting the atmosphere only: some typical values for this and various other combinations are shown in Table 1.

With all processes in the atmosphere the oceans act only as a source of thermal inertia, preventing the atmosphere developing the large oscillations that its small heat capacity would otherwise produce, and itself ending up with much the same behaviour as that of the 1-layer model, while the atmosphere reacts more and sooner (though not nearly as much as it would uncoupled). Annual mean temperatures for the two layers are thus equal, but the amplitude and phase of the seasonal cycle differ. As for the 1-layer case, seasonal forcing gives a net warming, which is shown in Table 1 (cf Fig 6). An almost identical warming (not shown) was produced by introducing the seasonal cycle into the other cases described below.

If all the radiation and diffusion are put in the ocean, again in theory the atmosphere, as the passive coupled part, should exhibit a smaller, lagged cycle, but as its heat capacity is so much smaller the effect is negligible (0.2% reduction in amplitude, and $2^{2}/_{3}$ day lag, according to analysis, contrasted with 42% and 54 days the other way around) and the values for this case are almost the same for the two layers, and very like the 1-layer case. With diffusion in the ocean and radiative processes in the air, as one would expect the air temperature has stronger meridional variation, while again having a seasonal cycle about twice that of the oceans. The reverse, all diffusion in the atmosphere and radiation in the oceans, naturally gave SST more meridional variation, and seasonal cycles of almost equal size. Finally the model was run splitting the quantities more or less as they are observed (values are taken from Fig 3.2 of NAS, 1975), with the surface absorbing most of the solar radiation but the atmosphere emitting most of the long-wave, and roughly the same diffusivity in each (Table 2). These proportions were also used for all runs of the 3-level versions of the model described in section c, except where otherwise indicated. They gave an atmosphere with a slightly

wider seasonal range than the oceanic layer, but about 2K colder (Table 1 and Fig 9). This, perhaps, suggests the linear exchange coefficient underestimates convective instability, and should increase with (T_m-T_a) for a more "realistic" simulation.

The basic rate of convergence is not much affected by these alterations: as the coupling between the 2 layers is fast compared to diffusion and heat loss to space, the latter dominate. Analytically the harmonics decay independently, the e-folding time for the n^{th} being given by the total heat capacity divided by the factor $(B + n(n+1)D)$ which includes the effects of decay by radiation to space and by diffusion, the latter becoming more important for higher harmonics or smaller scales. The comparative closeness to equilibrium of the two layers is, however, altered. For example in the case when all diffusive and radiative processes are in the atmosphere, and so the ocean acts only to delay and reduce temperature changes, the ocean will lag behind by a factor depending on the harmonic concerned: this is illustrated in Fig 10 for constant forcing. A very rapid adjustment as the atmosphere begins to react as if it were only damped by its own heat capacity, and then is slowed once it diverges from the SST, is followed by almost parallel convergence. Figs 11 and 12 show convergence towards the seasonal cycle (after a 30 year run with annual mean forcing) of SST and air temperature at low and high latitudes for the same case. Convergence is very similar for the other cases (not shown).

c. 3-layer versions

The introduction of a third layer introduces surprisingly many problems: how is its behaviour to be parametrized, and what, in fact, do we want to tune to? Fig 13 shows climatological zonal-mean oceanic temperatures for the Northern Hemisphere for various depths corresponding to the third layer of the 3-layer versions of the model. Plainly the layer is not vertically homogeneous - nor should we expect it to be: the oceanic mixed layer's name indicates it has this useful property, but in general stratification will be important. Towards the poles the Arctic halocline allows the existence of an apparently convectively unstable state and over two latitude belts the ocean actually warms towards the poles, not something that can be accounted for in any simple diffusive model. It certainly looks nothing like the 2-mode harmonic we would expect to fit the output of the model (solid curve shows one), and which is the best we can hope to tune to in a simple model omitting the effects of factors such as detail in the vertical, salinity and western boundary currents.

There are some simple constraints on this tuning. One version of the 3-layer model generalises the diffusive treatment of the 2-layer model, having a constant diffusivity in the deeper layer, a constant coefficient for heat exchange between it and the OML, and convective overturning when it is warmer than the OML at the same grid-box, essential to give it a lower mean temperature.

The equations for the three layers are then:

$$C_a \frac{dTa}{dt} + \frac{d}{dx}\left\{(1-x^2).DO.r_3 \frac{dTa}{dx}\right\}$$

$$= \lambda_{am}(T_m-T_a) + r_1S(x,t) - r_2(A + BT_a)$$

$$C_m \frac{dTm}{dt} + \frac{d}{dx}\left\{(1-x^2).DO.(1-r_3) \frac{dTm}{dx}\right\}$$

$$= \lambda_{am}(T_a-T_m) + \lambda_{md}(Td-Tm) + (1-r_1)S(x,t)-(1-r_2)(A+BT_m)$$

$$Cd \frac{dTd}{dt} + \frac{d}{dx}\left\{(1-x^2).DD.\frac{dTd}{dx}\right\}$$

$$= \lambda_{md}(T_m-T_d)$$

with the condition that if $T_m<T_d$, both are set to $\frac{CmTm + CdTd}{Cm + Cd}$

where the suffixes a, m and d refer to the atmosphere, OML and intermediate waters respectively, r_1, r_2 and r_3 are the proportions of incoming radiation, outgoing longwave and diffusion respectively to affect the atmosphere (given in Table 2), and $S(x,t)$ is the prescribed insolation absorbed. In this case to get a reasonable description of the time evolution of the system the exchange coefficient λ_{md} must be near the correct value. Then to get an approximately correct temperature in the tropical third layer, which can be thought of as being produced by a balance between diffusion of colder intermediate water from the poles and warmer water from above (see right-hand graphs of Fig 14 - this figure will be discussed more fully later) the lower layer diffusivity is fixed. However, combined with that already present in the upper layers this gives too much lateral diffusivity overall and distorts the SST and air temperatures. Fig 15 compares results with various values of the relevant parameters. (These were not run to full equilibrium, and are not entirely comparable). In all cases, the general pattern is of the parabola of the 2-mode approximation, flattened off towards the poles by the effects of convective overturning. (This process warms the OML and removes the rather awkward situation of a nominal ocean temperature of -8°C at the cost of an annual mean air temperature which remains positive everywhere. Results, not shown, from runs including a simple treatment of the insulating effect of sea-ice show that this problem does not then occur). When the lower layer diffusivity (DD) was reduced the lower layer/OML coupling constant (λ_{md}) was reduced in the same proportion. The latter change cools the lower layer overall, as it makes the linearized exchange with the OML, which can only warm the lower layer, weaker in comparison with convective overturning, which can only cool it. However, the reduction in diffusivity increases the meridional variation of lower layer temperature, and these two effects tend to compensate so that its low latitude values, as intended, change little. Both alterations tend to increase the range of SST, but the mean value changes little, as it is constrained by the radiation balance.

Fig 15 also shows results from reducing the diffusivity in the upper layers, as suggested by an approximate calculation of the model's meridional fluxes which implied that the heat transport in the intermediate waters must be of the same order as that in the OML and atmosphere combined, to reach a reasonably well-tuned result. Thanks to the gross simplifications, this does

not of course imply that this is true of the real situation, or of ocean/atmosphere GCMs, as the tropical intermediate waters are cooled by effects such as upwelling rather than just horizontal diffusion. As can be seen, this produced quite realistic SST and intermediate water temperatures (and also brought mean air temperature at the "polar" grid box down to -4°C - not shown), though suggesting that the coupling between the two layers should be stronger. Fig 16 shows equilibrium results from a run with recommended parameter values for this version of the model, DO = 0.35, DD = $1 Wm^{-2}K^{-1}$ and $\lambda_{md} = 1 Wm^{-2}K^{-1}$. It may seem disappointing that we end up with so low a value of λ_{md} - Dickinson (1981) took the range 1-4 $Wm^{-2}K^{-1}$ as corresponding to observation: however this was an overall value whose largest contribution comes from convective overturning, which is here treated separately - if we replaced this by a variable λ_{md} increasing towards the poles, the tuned mean value would be higher.

Fig 17 shows convergence to this equilibrium. At the equatorial and mid-latitude grid boxes, the SST converges much as before (cf Figs 10 and 11) with the intermediate water temperature equilibrating more slowly and hardly changing the SST in the process. This is rather unlike the behaviour in Dickinson's and Roberts's models, where the OML settles down quite fast into a quasi-equilibrium about halfway between the intermediate water temperature and its final equilibrium, and then both layers converge together more slowly. The difference is partly due to the smaller value of λ_{md}, and partly due to the importance of the higher harmonics, which,decaying faster than global mean temperature, are even less affected by coupling between the layers. Both of these effects decrease the influence on the OML of the lower layer relative to that of radiative and diffusive decay in the upper two layers. At the pole, however, where convective overturning gives a close coupling between the two oceanic layers, they evolve almost in parallel, and actually both overshoot together. This is presumably due to the strong coupling between the two layers allowing them both to converge towards the equilibrium for the 2-layer case before the slower acting lower layer diffusion counteracts this. Similar temperature overshooting has been observed in the global mean temperatures in spinup of the GFDL ocean model (Bryan et al, 1975) which is almost identical to the Meteorological Office OGCM and has been hypothesized to be a general feature of thermohaline adjustment throughout the oceans (Rooth, 1982); it cannot be produced in a purely linear model such as that of Dickinson or Roberts. Atmospheric temperature (shown only for the equatorial point) is everywhere in quasi-equilibrium with the OML, as in Dickinson's and Roberts's models.

Rather more tunable parameters are available when the simple diffusive treatment is replaced, as far as the ocean is concerned, by an imposed meridional circulation, which is made possible by the presence of the third layer. The diffusivities for the ocean layers are replaced by mass fluxes between each of 19 pairs of adjacent boxes (which, by continuity, must be equal and opposite in the two layers, and which determine the up- or down-welling). A first-guess field was calculated by symmetrizing the zonal mean net heat fluxes down into the ocean given in Table 2 of Esbensen and Kushnir (1981) and deriving the mass fluxes which would provide the same meridional heat fluxes in the OML alone for the annual mean temperature field of the 2 level model. This neglected completely the meridional heat fluxes in the intermediate waters. Also, the effective oceanic diffusivity produced was much greater at low latitudes, making it impossible to find a value for the

atmospheric diffusivity, still kept constant, which gave a realistic temperature field. The values were therefore empirically tuned by reduction at low latitudes and increase at higher latitudes until a satisfactory equilibrium was given. The circulation was kept the same in both hemispheres, though of course it would have been possible to try to simulate the effects of their different real circulation patterns.

Table 3 shows the final values, and Fig 18 shows the results, compared with those for circulations 0.25 and 2.5 times as strong as that decided on. The proportions of solar and longwave radiation to affect the atmosphere were set to 0.58 and 0.55 respectively for these runs, rather than the values 0.37 and 0.76 generally used, to avoid the problem of the atmosphere being generally about 2 K cooler than the OML referred to above. This gave a more realistic simulation in the tropics, but worsened the already over-cold poles. An atmospheric diffusivity of 0.549 Wm^{-2} K^{-1} was used for these and all other runs of the circulating model shown.

The chosen strength of circulation gives zonal-mean SST equatorwards of about 17° almost constant at 27°C, as observed (Levitus, 1982) which is gratifying, even if fortuitous, and a "knee" in the intermediate water temperature at about 50°, reminiscent of that in Fig 13 (though the latter is further polewards, and too small to be distinguishable on this grid). Polar temperatures are again too low in both oceanic layers, since the model does not allow for the insulating effect of sea ice. Further tuning could no doubt be carried out if it was desired to simulate more closely the observed temperature distribution, but this seems quite good enough for our purposes. Increasing the strength of the circulation cools the low latitude OML and warms it at high latitudes, as one would expect; it warms the high latitude intermediate waters, strongly coupled by convective overturning, and tends to reduce the zonal gradient in the lower layer; these effects counteract each other to give a smaller and inconsistent effect on tropical intermediate water temperature.

Although, as mentioned on page 3, the total meridional fluxes of an EBM will be insensitive to its details, it is interesting to compare the contributions of the various processes to the local heating for the two versions of the 3-level model. We should expect the differences to be mostly in the oceans, and in fact the gross features of the heat balance for the OML and atmosphere (Fig 19) are the same. The dominant terms are the prescribed insolation absorbed, the infrared emitted, weakly dependent on temperature, and the heat loss from sea to air. Although the large value of λ_{md} (45 $Wm^{-2}K^{-1}$) makes the latter apparently strongly temperature dependent, it varies less than might be expected between the two models as the air-sea temperature difference adjusts so as to make it account for most of the local radiative imbalance. The residual of these three terms is therefore rather smaller, and the largest of the horizontal transport terms accounting for it is the atmospheric diffusion, particularly in the circulating model with its higher atmospheric diffusivity. A noticeable difference in the oceanic fluxes (Fig 14) is the far weaker convective overturning in the circulating model, where the lower layer at high latitudes is not warmed by diffusion, and so is only affected by mass flux from above which, in a non-seasonal equilibrium, would lead to no convective overturning at all. Other differences are that in the diffusive model the "horizontal" fluxes in each layer must sum to zero while in the circulating model they are all of the same sign, and that the "vertical" flux is limited

in occurrence in the circulating model. There is also a conspicuous difference in the sea-air flux and compensating changes in the atmospheric diffusive heating (Fig 19). The falling off of the former towards the pole is much more marked in the circulating model, where more of the meridional transport is in the atmosphere, and is not significantly affected by the much weaker convective overturning. The sea-air flux is, however, reduced in the tropics where the OML is cooled by upwelling, giving an overall pattern quite like that given by e.g. Fig 1 of Esbensen and Kushnir (1981).

Fig 20 shows convergence rates with the standard circulation (but the same radiative parameters as Fig 17, for consistency, so that the equilibrium is warmer than Fig 18). The pattern is more complex than Fig 17, although again convergence is much faster for the OML than the lower layer except at the poles where both converge at an intermediate rate. Elsewhere, however, there is now a delay before convergence of the lower layer begins due to the need for the box feeding the box concerned to adjust: one of the boxes shown takes about 30 years to start while another needs a century. These also cause small irregularities in the behaviour of the OML above. The absence of overshooting at the poles suggests a greater coherence in the convergence of the two layers for the circulating case due to the closer coupling between the two layers at any one point, and the absence of distinct timescales associated with different mixing processes.

4. Summary

A time-dependent zonal-mean energy balance model has been written which is intended to be suitable for studying the effects of accelerated convergence schemes necessary for more complex and computationally expensive coupled ocean-atmosphere models and perhaps for gaining insight into the reaction of the ocean to changes in atmospheric fluxes more generally. The emphasis in writing the 1-layer version, which acted as a test-bed for the numerical techniques, and the 2-layer version, which allows in a very simple way for the interaction of atmosphere and ocean, has been on the simplicity of the parametrizations used. The model's behaviour then depends only on the values of certain parameters which all have a simple physical interpretation, and altering which gives an intuitively reasonable result. The linearity also makes it possible to treat the model analytically. However, the coding is intended to make it easy to introduce more complex "interactive" parametrizations of such variables as albedo and meridional heat transports, which might have a significant effect on the convergence of the model.

The two versions of the 3-layer model include an additional, non-linear, process, convective overturning on a simple temperature criterion, which makes their behaviour non-analytic, although still qualitatively similar. The extra heat capacity of the weakly coupled lower layer considerably delays reaching final equilibrium, and causes convergence to occur on multiple timescales. This makes their behaviour more like that of the OGCMs it is intended to simulate. They remain, of course, much cruder in formulation, and lacking in feedbacks which are present in the real system or in coupled GCMs, but which cannot be represented properly in such a simplified model. It is hoped, however, that they contain enough of the essential processes that asynchronous coupling schemes will cause similar perturbations in both as in a full OGCM.

Table 1. Annual mean and seasonal range of SST and air temperature at a high, middle and low latitude belt in the 2-layer model with various parameter values, and with seasonal forcing unless otherwise indicated.

	Annual Mean			Seasonal range		
	Pole (65°-90°)	30°-37°	Equator (0°-6°)	Pole (65°-90o)	30°-37°	Equator (0°-6°)
All radiative and transport processes in the atmosphere, annual mean forcing						
SST	-8.21	15.98	29.37	0.00	0.00	0.00
Air	-8.21	15.98	29.37	0.00	0.00	0.00
All radiative and transport processes in the atmosphere, seasonal forcing						
SST	-5.79	17.44	30.36	4.16	3.22	0.33
Air	-5.79	17.44	30.36	7.11	5.51	0.57
All radiative and transport processes in the OML						
SST	-5.79	17.45	30.36	4.36	3.50	0.36
Air	-5.79	17.45	30.36	4.35	3.50	0.36
Radiation in the atmosphere and transport in the OML						
SST	-5.14	17.42	29.95	4.18	3.35	0.35
Air	-6.89	17.42	31.18	7.13	5.83	0.61
Transport in the atmosphere and radiation in the OML						
SST	-6.89	17.42	31.18	5.14	3.52	0.37
Air	-5.14	17.42	29.95	4.34	3.35	0.34
37% of solar radiation, 76% of longwave radiation and 40% of diffusion occurring in the atmosphere						
SST	-4.47	19.01	32.06	4.31	3.45	0.36
Air	-6.07	16.95	29.74	4.75	3.78	0.39

Table 2. Values used in the various versions of the model for certain parameters. (All units $Wm^{-2}\ K^{-1}$)

	Diffusivity of atmosphere and OML combined DO	Lower layer diffusivity DD	Lower layer/OML heat exchange coefficient λ_{md}
1-layer model	0.649	-	-
2-layer model	0.649	-	-
3-layer diffusive model	0.35 (except as indicated in caption to Fig 15)	1	1
3-layer circulating	0.549 (all in the atmosphere)	0	0

The OML/atmosphere heat exchange coefficient was 45 $Wm^{-2}\ K^{-1}$ throughout. Except where otherwise indicated, the atmosphere had 60% of DO, 37% of incoming solar radiation, and 76% of outgoing infrared. The 1-layer model was normally run with 48 or 60 timesteps to the year, which, as mentioned in section 3a, is few enough to give a detectable truncation error. The 2 and 3-layer models were run with 360 or occasionally 120 timesteps to the year. Like the Dynamical Climatology Branch's atmospheric GCM, the Budyko-Sellers model has a calendar with 360 days to the year.

Table 3. Circulation prescribed for the 3-layer circulating model.

Polewards flow in the OML from the i^{th} box away from the equator to the $(i+1)^{th}$, in fractions of the total contents per year.

1	2	3	4	5	6	7	8	9
2.354	3.812	2.978	2.537	2.432	2.407	2.116	1.491	0.912

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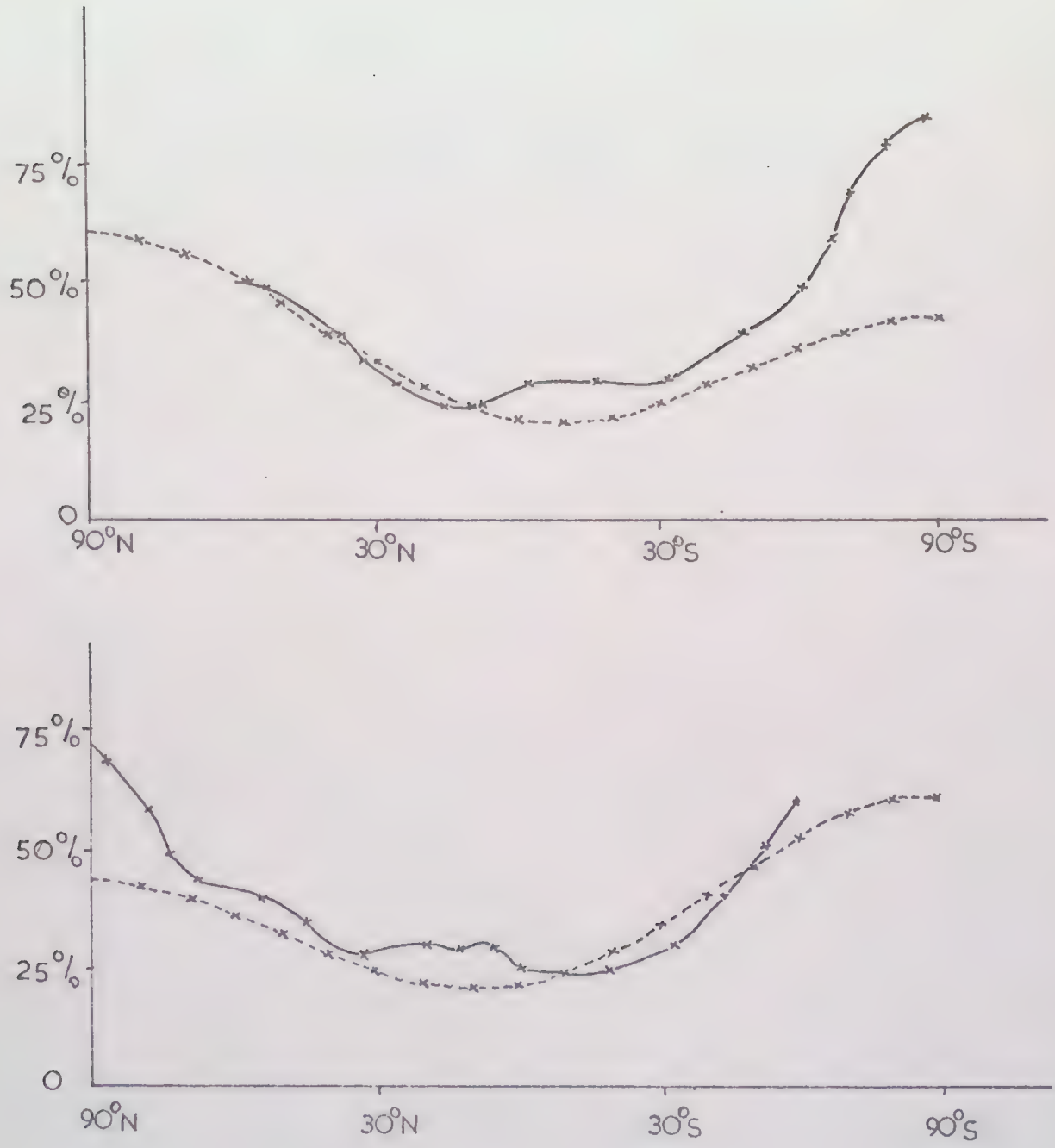
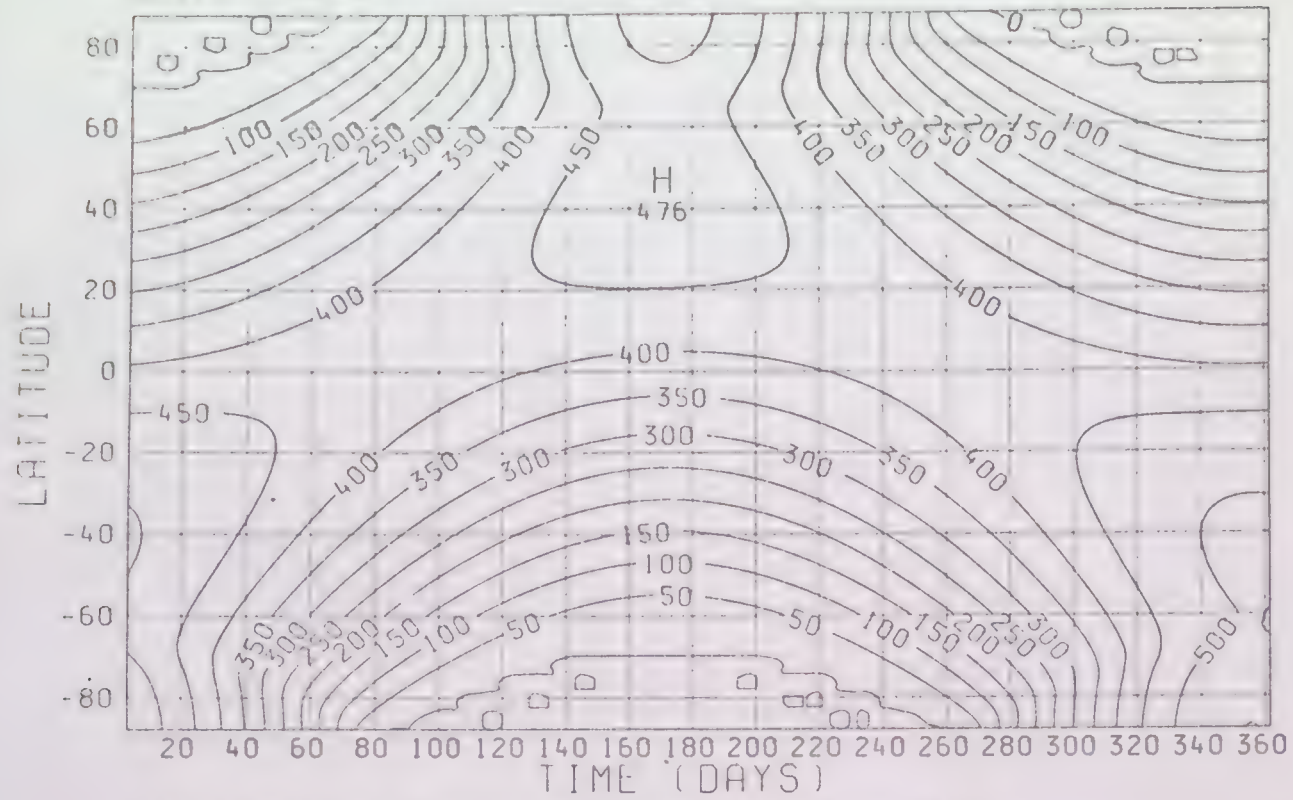


Figure 1. Zonal mean albedo as a function of latitude at the end of January (upper graph) and July (lower graph) from NOAA satellite observations in 1975 and 1974 (solid curve) and as given by the second-mode approximation used in the model (dashed curve).

TIME LATITUDE DIAGRAM OF INCOMING INSOLATION
AS GIVEN BY THE ASTRONOMICALLY CORRECT FORMULA
BEGINNING AND ENDING AT THE DECEMBER SOLSTICE
UNITS - WM-2



TIME LATITUDE DIAGRAM OF INCOMING INSOLATION
AS GIVEN BY THE MODEL'S 2ND HARMONIC APPROXIMATION
BEGINNING AND ENDING AT THE DECEMBER SOLSTICE
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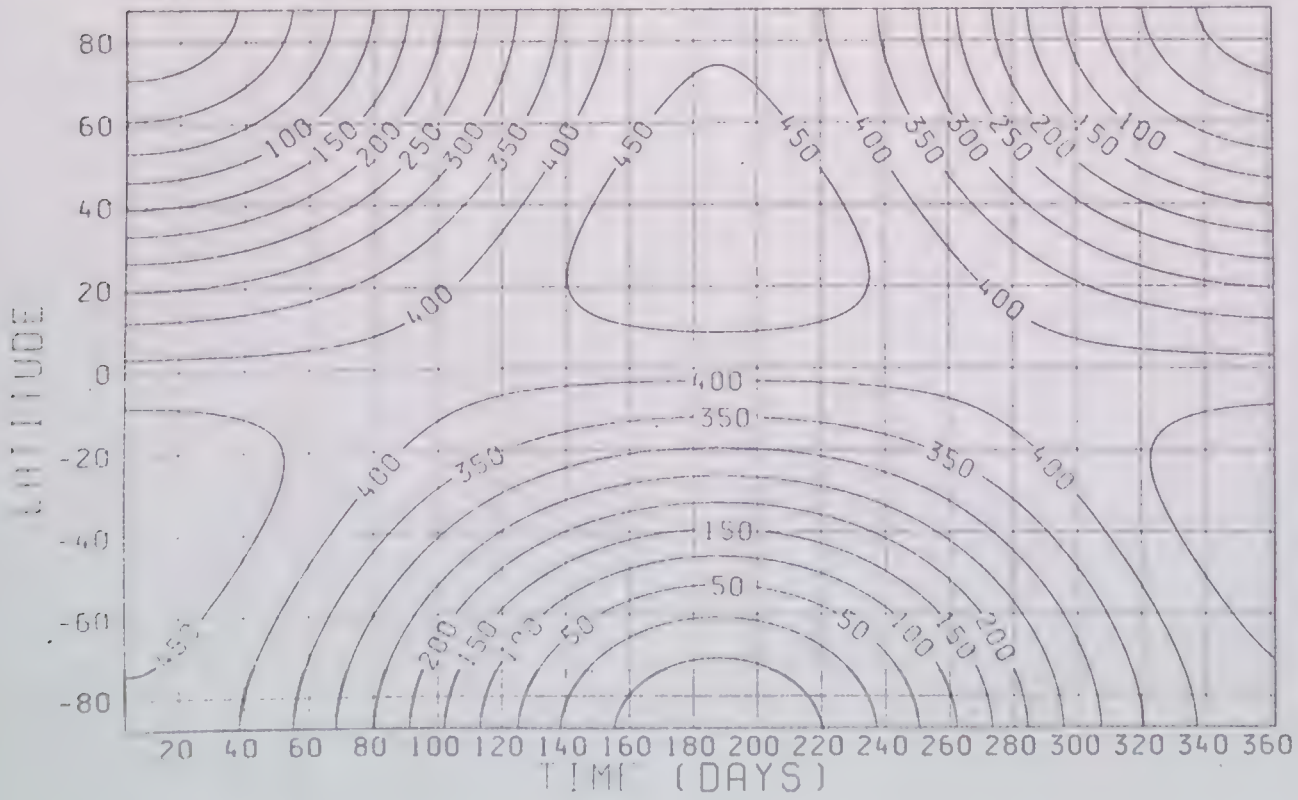

Figure 2.

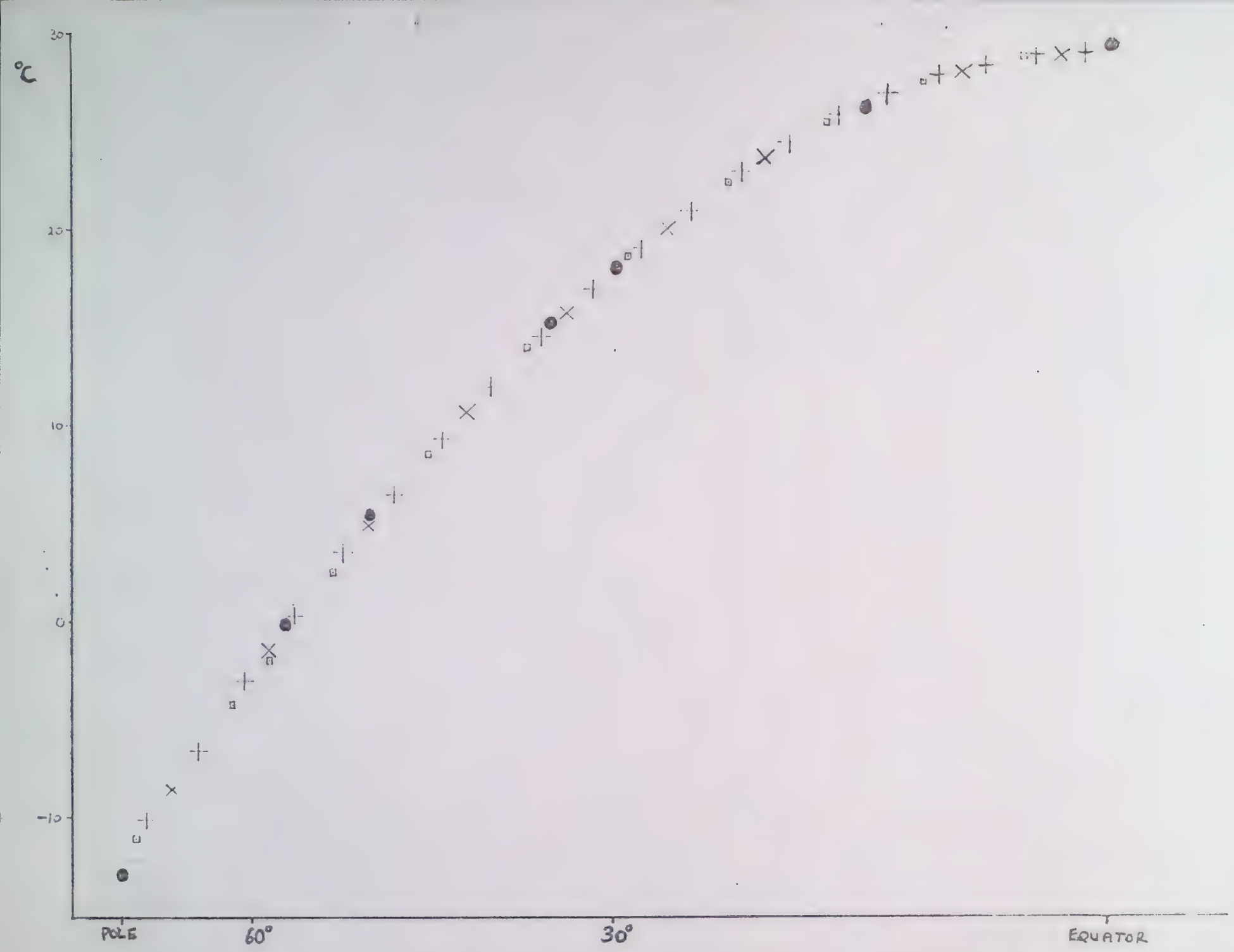

Figure 3. Equilibrium of the 1-layer model with annual mean forcing: analytic values (●) and numerically obtained ones for spacing of 10 (X), 20 (+) or 40 (□) latitude belts to a hemisphere.

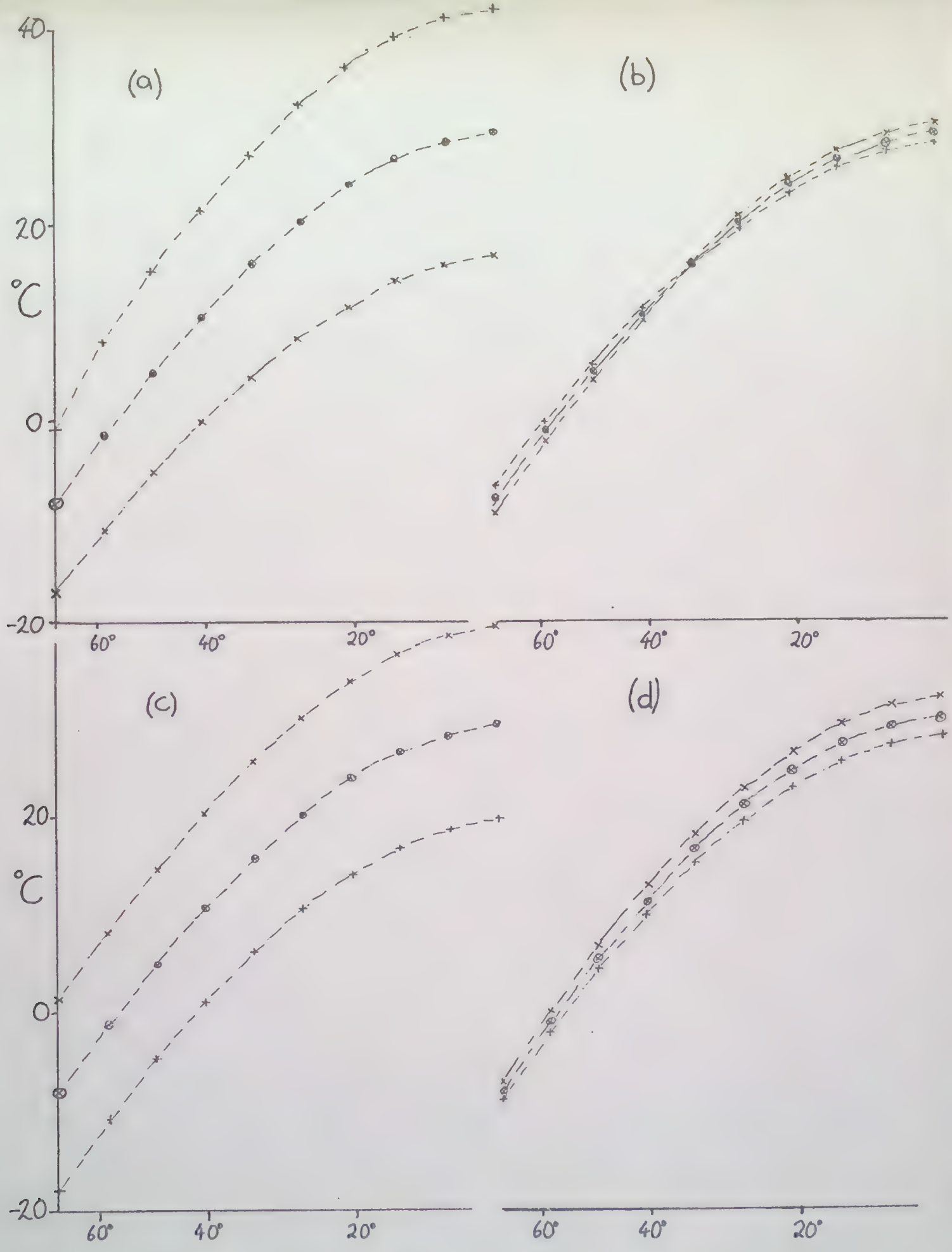

Figure 4. Sensitivity of the equilibrium of the 1-layer model with annual mean forcing to ± 10% change in parameter values:
(a) Solar constant, (b) Diffusivity, (c) A_2, (d) B.
0 represents standard values, + 10% larger x 10% smaller.
In all cases the change in global mean temperature agreed with the analytic result to at least 0.001 K, and the change in higher harmonics was qualitatively as predicted.

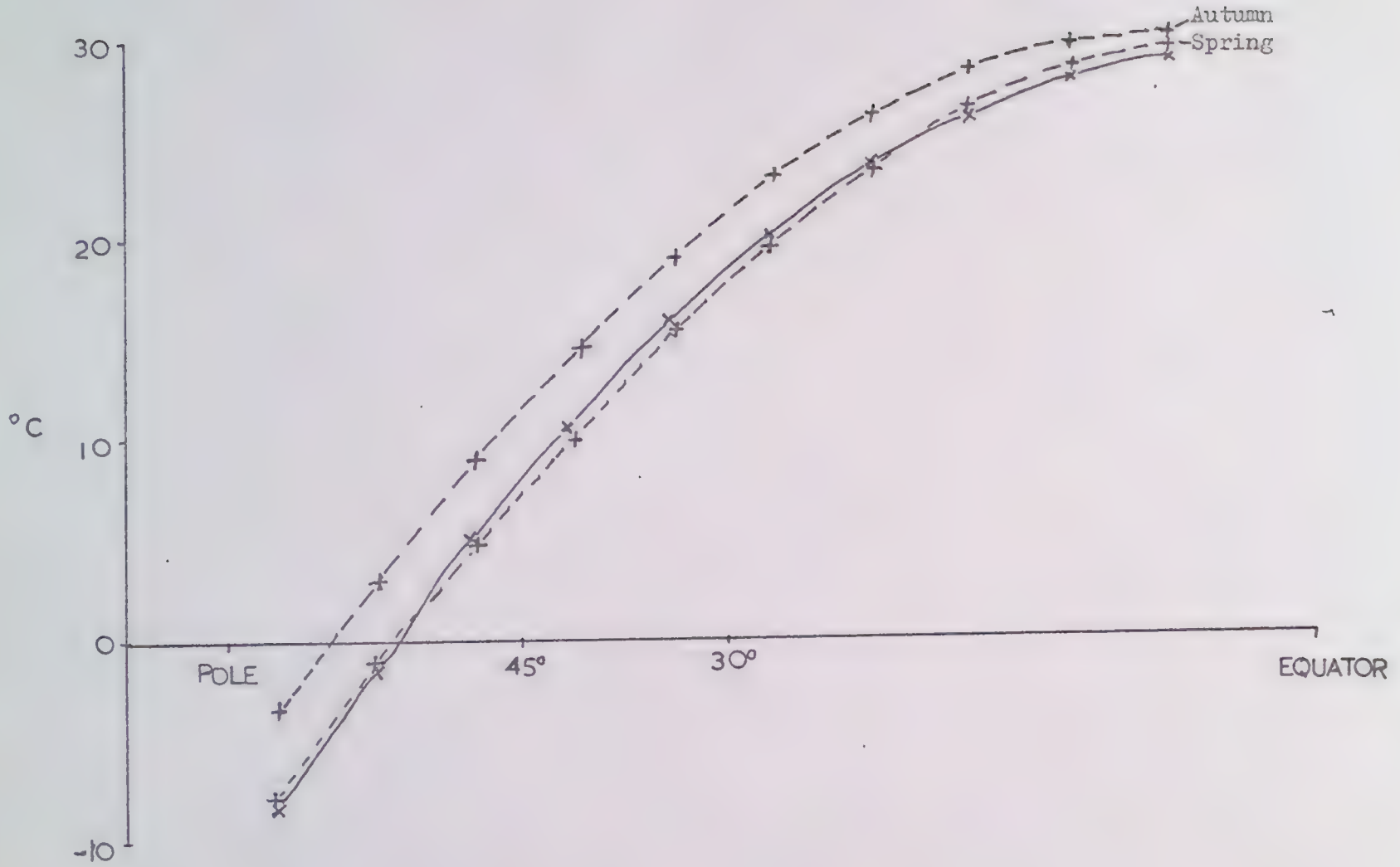

Figure 5. Equilibrium of the 1-layer model with annual mean forcing (X) compared with temperatures at the equinoxes in the equilibrium seasonal cycle when seasonally forced (+).

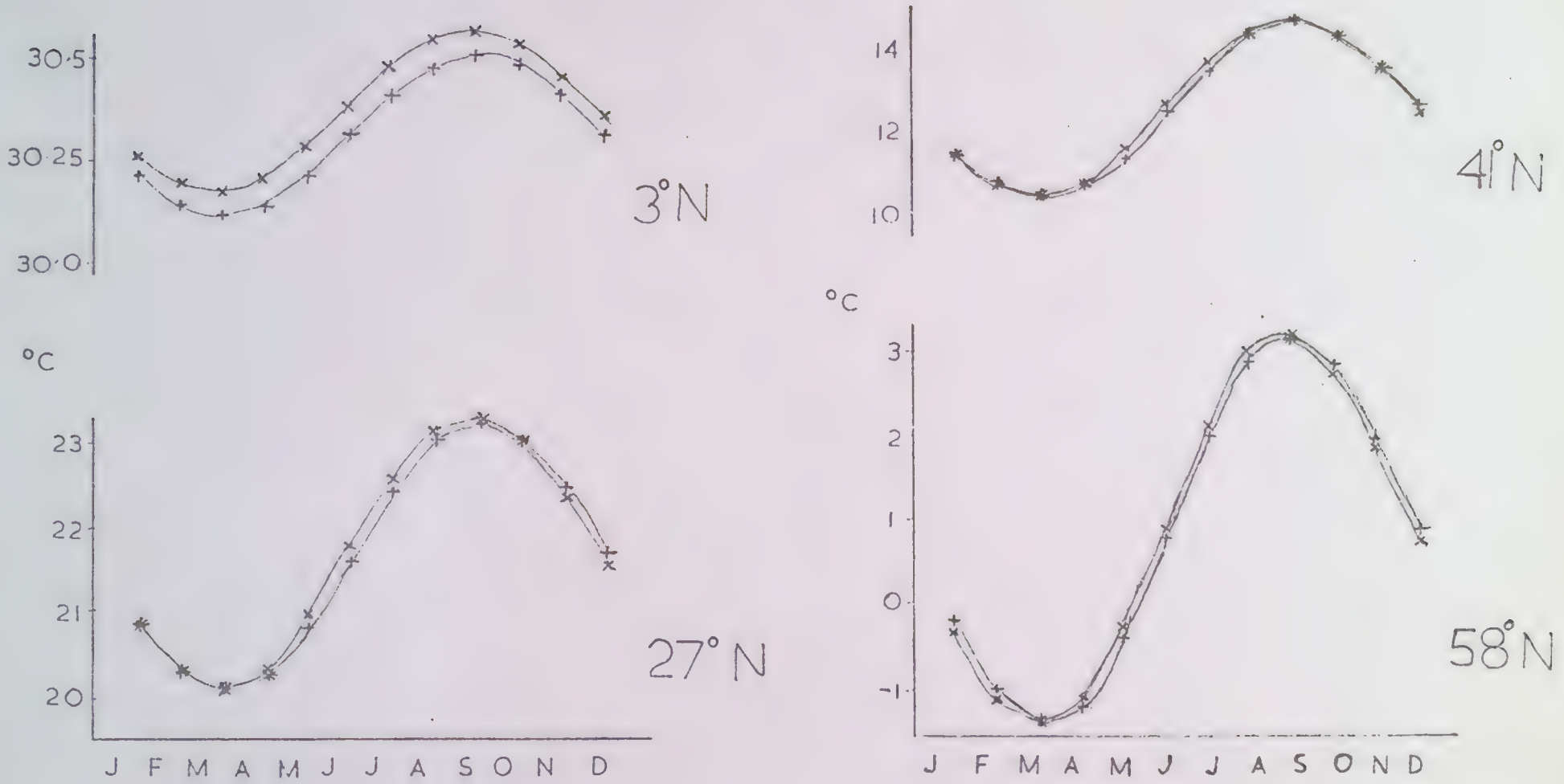

Figure 6. Seasonal cycle at four points in the 1-layer model after 50 years' integration with 15 points to a hemisphere and 48 timesteps to a year (×) compared with the analytic equilibrium seasonal cycle for corresponding latitudes (+). (Note temperature scales differ.)

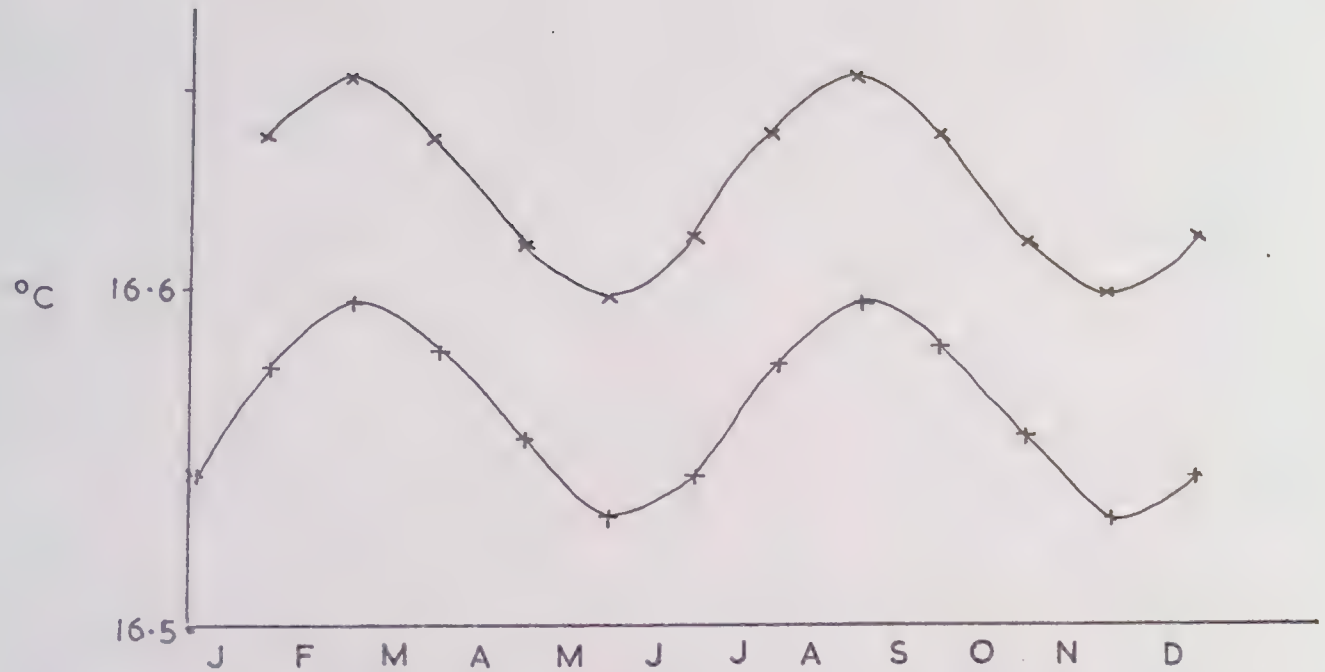

Figure 7. Equilibrium seasonal cycle of global mean temperature in the seasonally-forced 1-layer model: analytic values (+) and those obtained numerically after 30 years' integration (X) with 10 points per hemisphere and 48 timesteps a year.

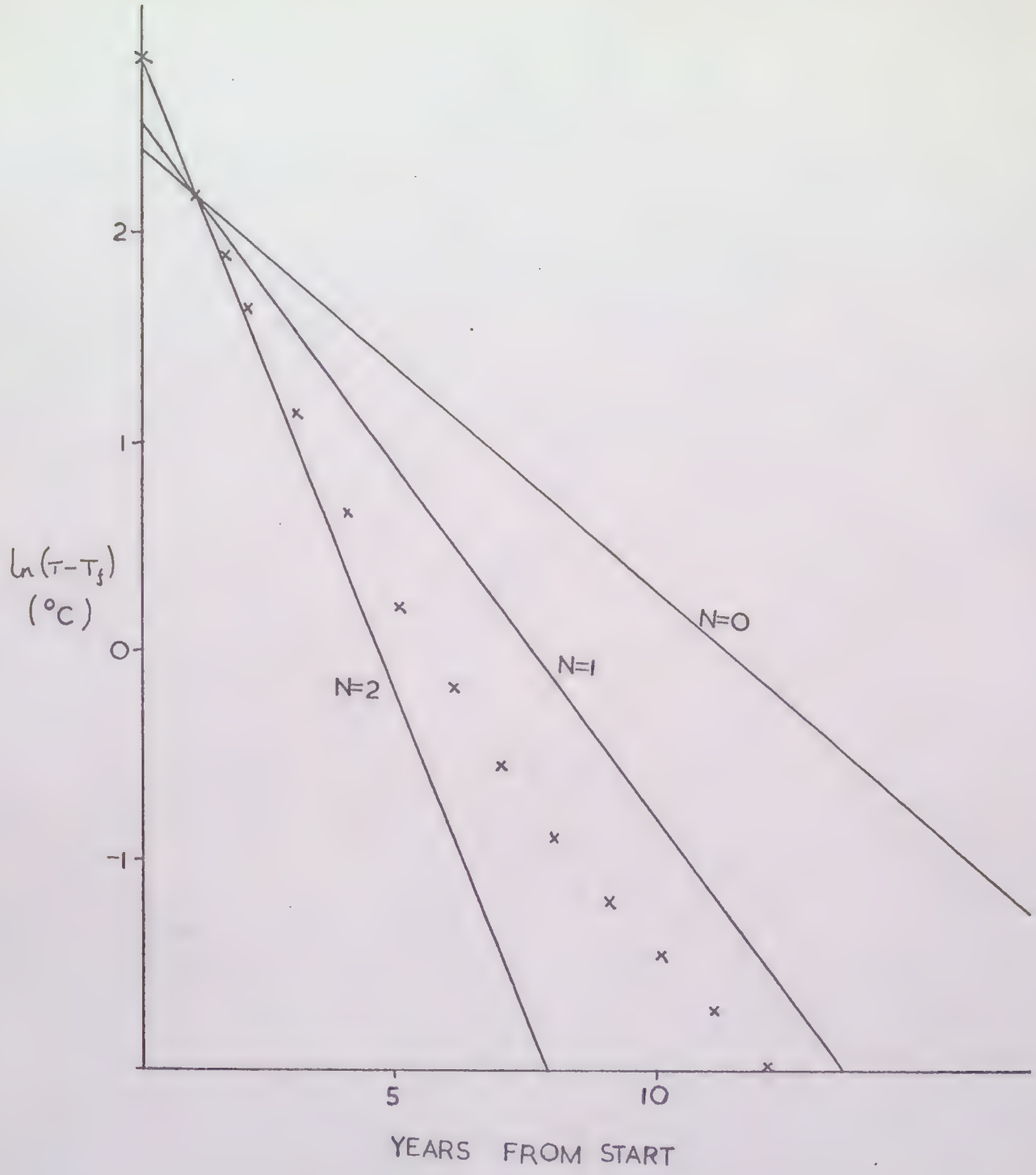

Figure 8. Convergence of end of year temperature at a low-latitude point (about 6°S-0°) in the seasonally-forced 1-layer model. (T_f is the value reached after 30 years). Straight lines show analytic decay rates for the first three harmonics.

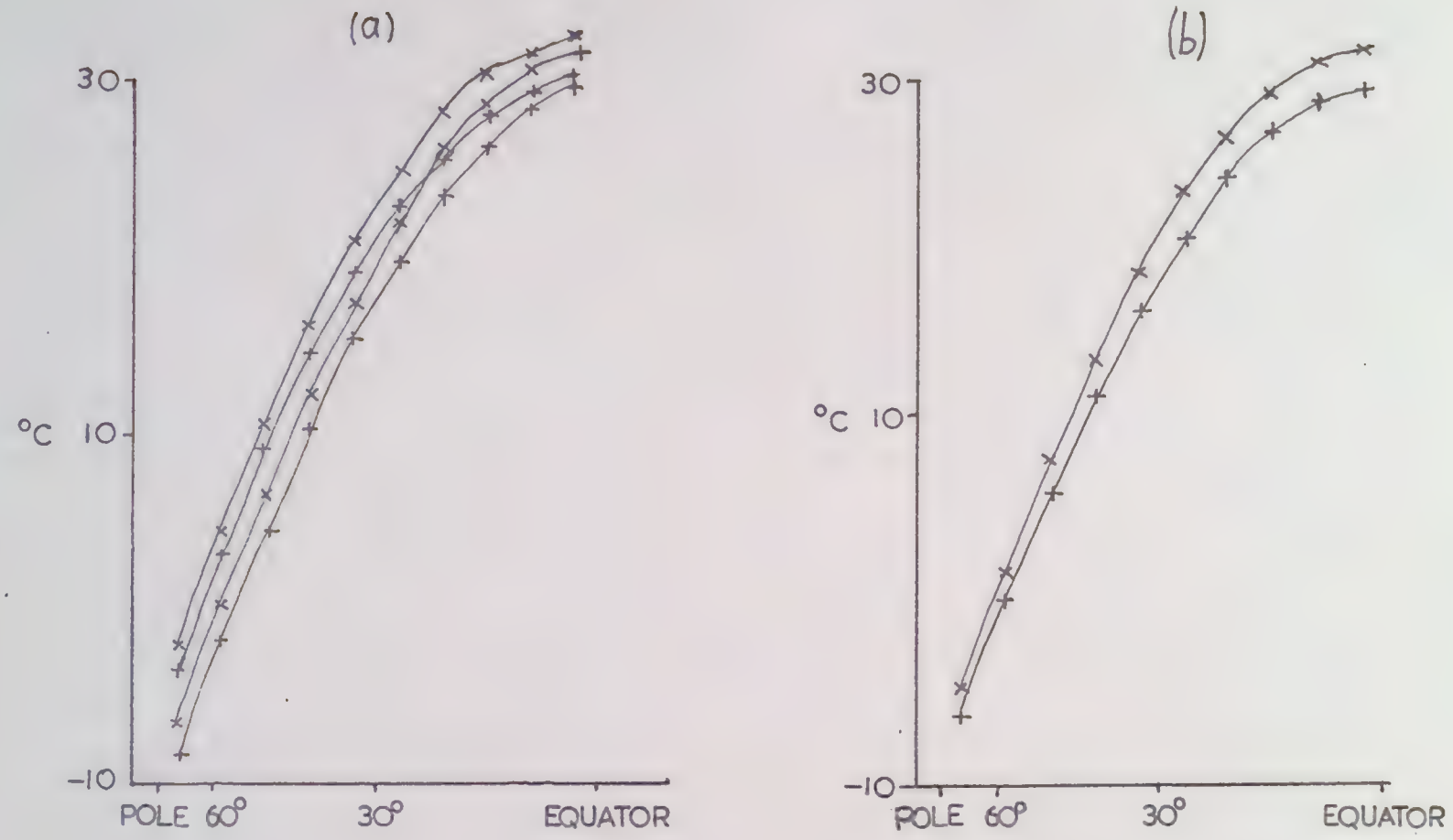

Figure 9. Results from the 2-layer model with 37% of absorbed insolation, 76% of longwave emission and 40% of diffusion affecting the atmosphere.
(a) Maximum and minimum SST(X) and air temperature (+) for the equilibrium seasonal cycle. (b) Equilibrium annual mean temperatures with seasonal forcing and with annual mean forcing.

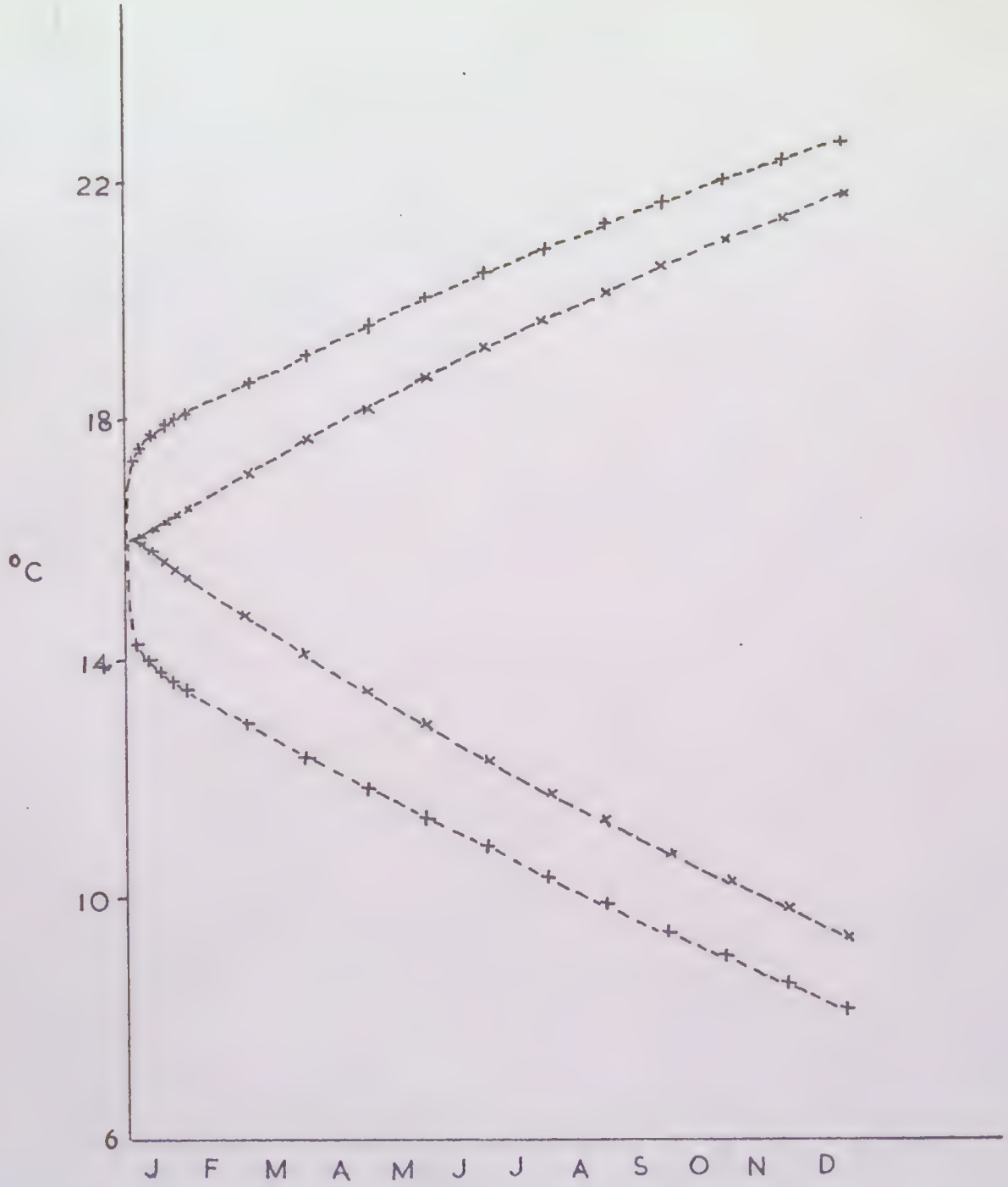

Figure 10. Convergence of SST(X) and air temperature (+) for a high latitude (53°-65°) and a low latitude (0°-6°) point through the first year of integration of the 2-layer model with annual mean forcing and all diffusive and radiative processes affecting the atmosphere, starting from a constant value of 16°C. Equilibrium values were -1.5°C, +29°C. As the second harmonic is dominant, the air temperature would be expected to be about $^{8}/_{9}$ as far from equilibrium as the SST.

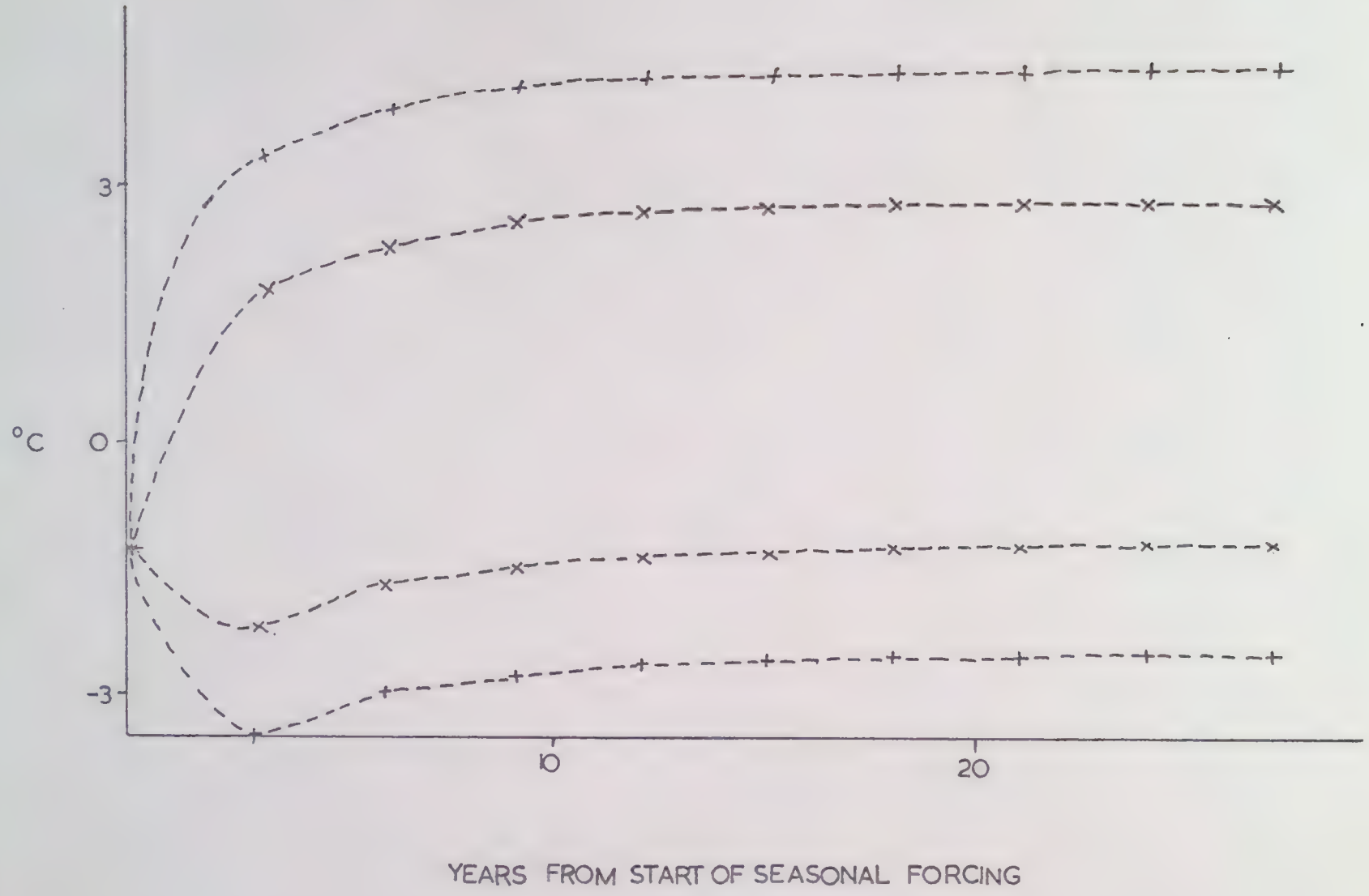

Figure 11. Variation of annual maximum and minimum of SST(X) and air temperature (+) at a high latitude point in the 2-layer model with all diffusive and radiative processes in the atmosphere, starting from the equilibrium with annual mean forcing.

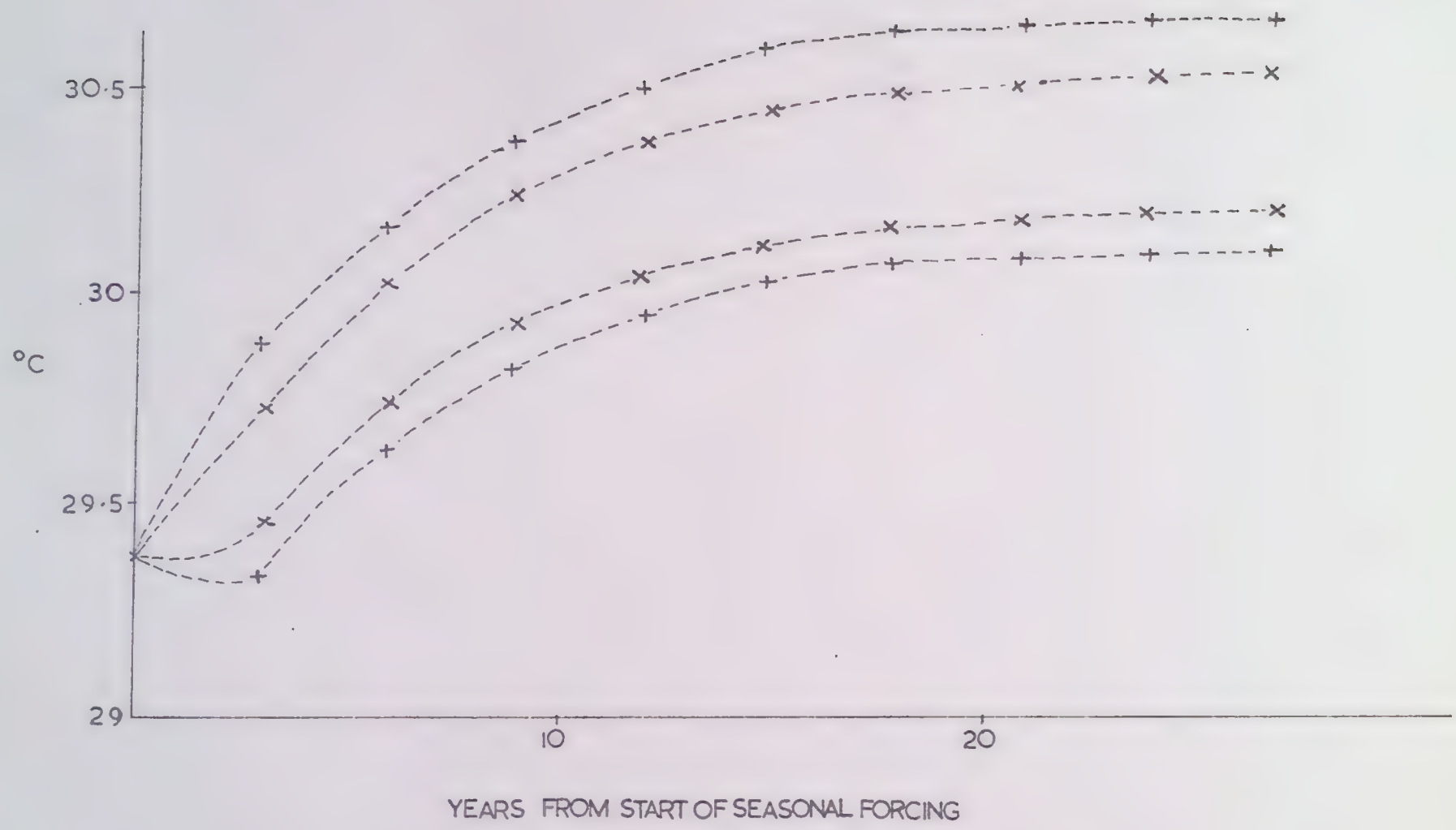

Figure 12. As Figure 11, but for a low latitude point (and with change of scale).

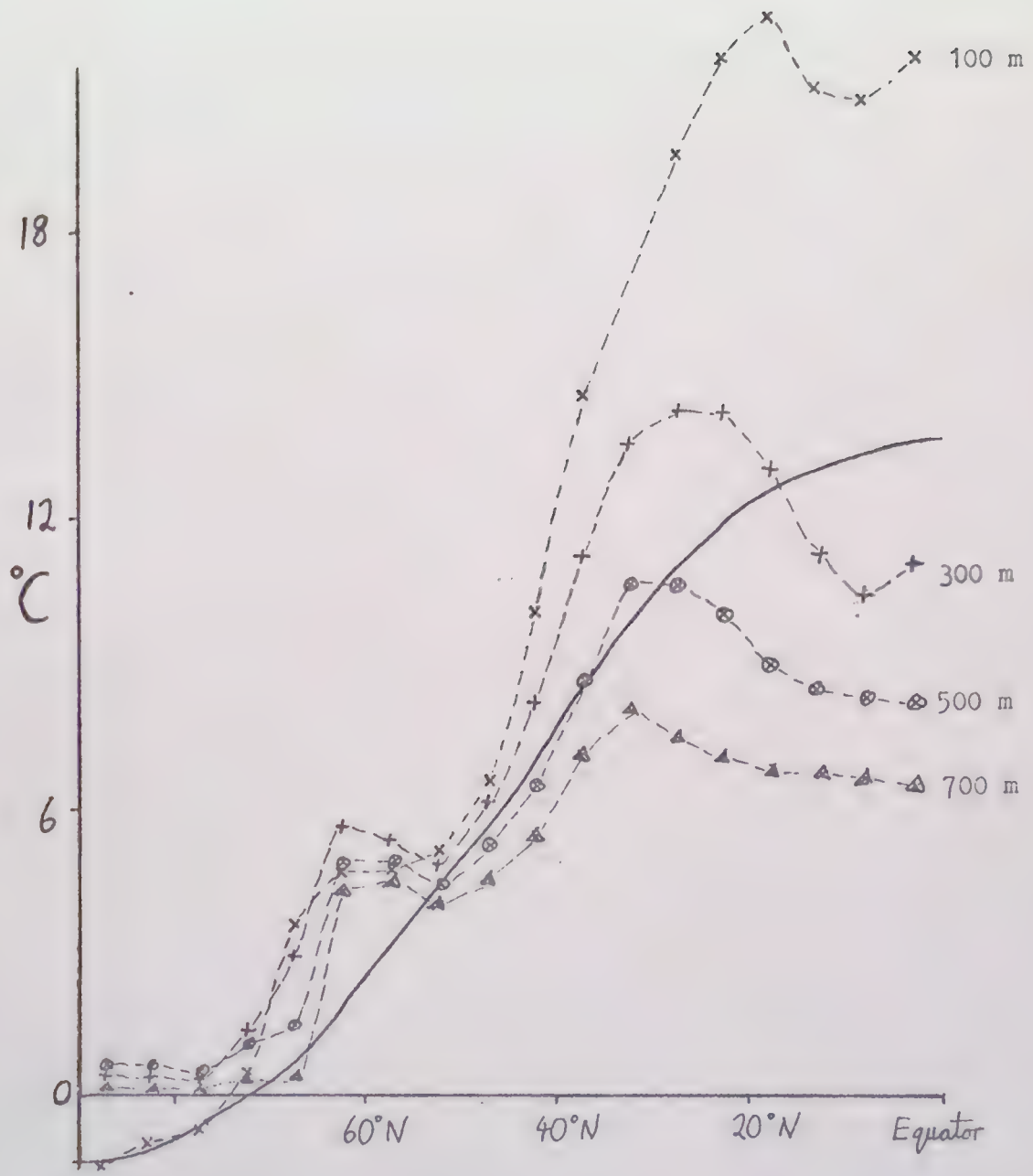

Figure 13. Zonal mean oceanic temperatures for the Northern hemisphere at 100m (x), 300m (+), 500m (⊗) and 700m (▲), from Levitus (1982), with a parabola in sin(latitude) for comparison. (Note abscissa is latitude in degrees.)

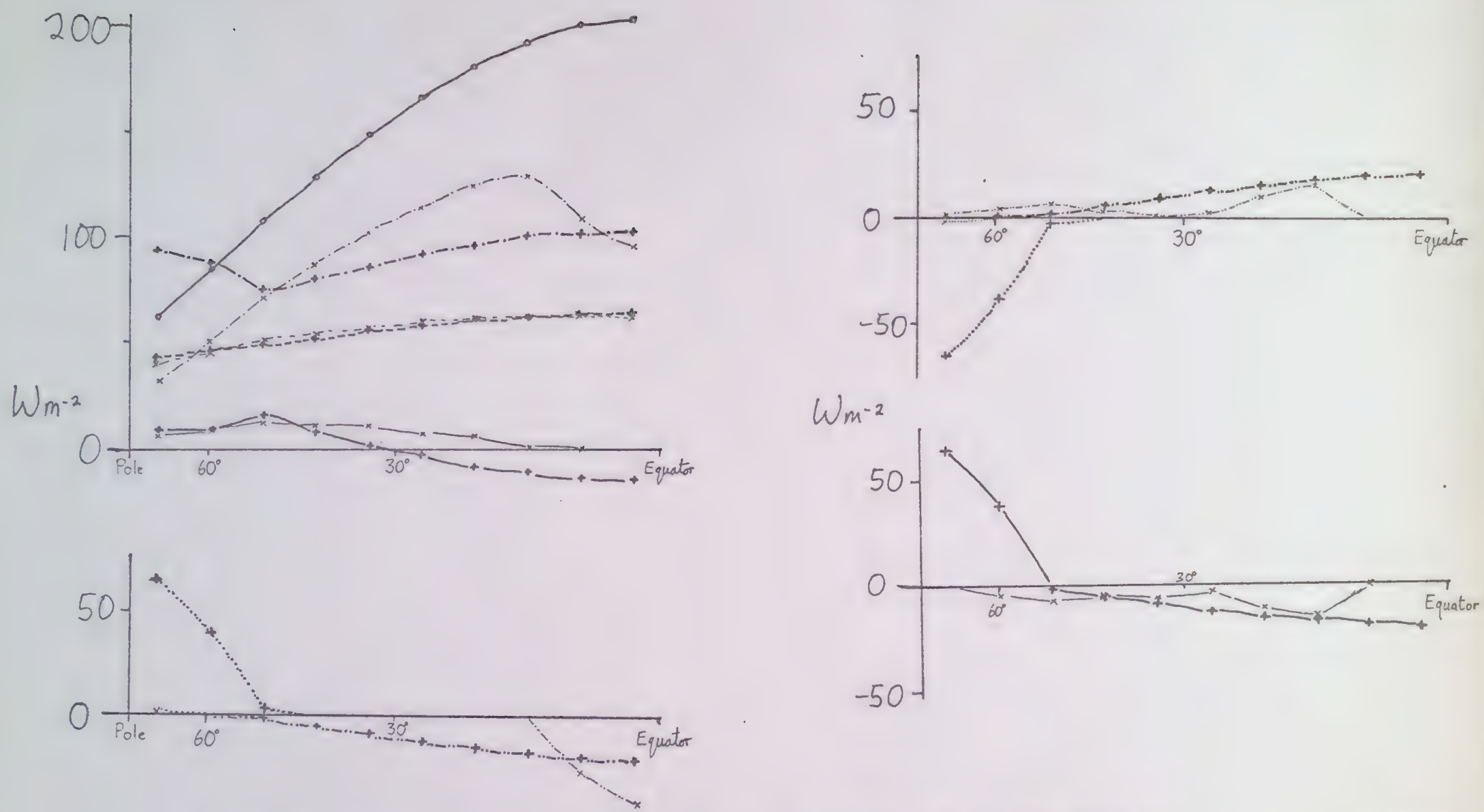

Figure 14. Annual mean heating produced by the various transport mechanisms in the OML (left) and intermediate water (right) of the all-diffusive (+) and circulating (x) versions of the 3-level model with standard parameter values, corresponding to Figs 16 and 18.
o——o insolation absorbed, x-·-·x sea-air exchange, x----x infrared emissions, x——x convergence of diffusive fluxes, or heating due to advection, x.....x flux due to convective overturning, x—·—·—x effects of exchange between OML and intermediate waters.

Note that on the left hand graph, heat loss to the atmosphere and by radiation are shown as positive: otherwise all signs are physically correct.

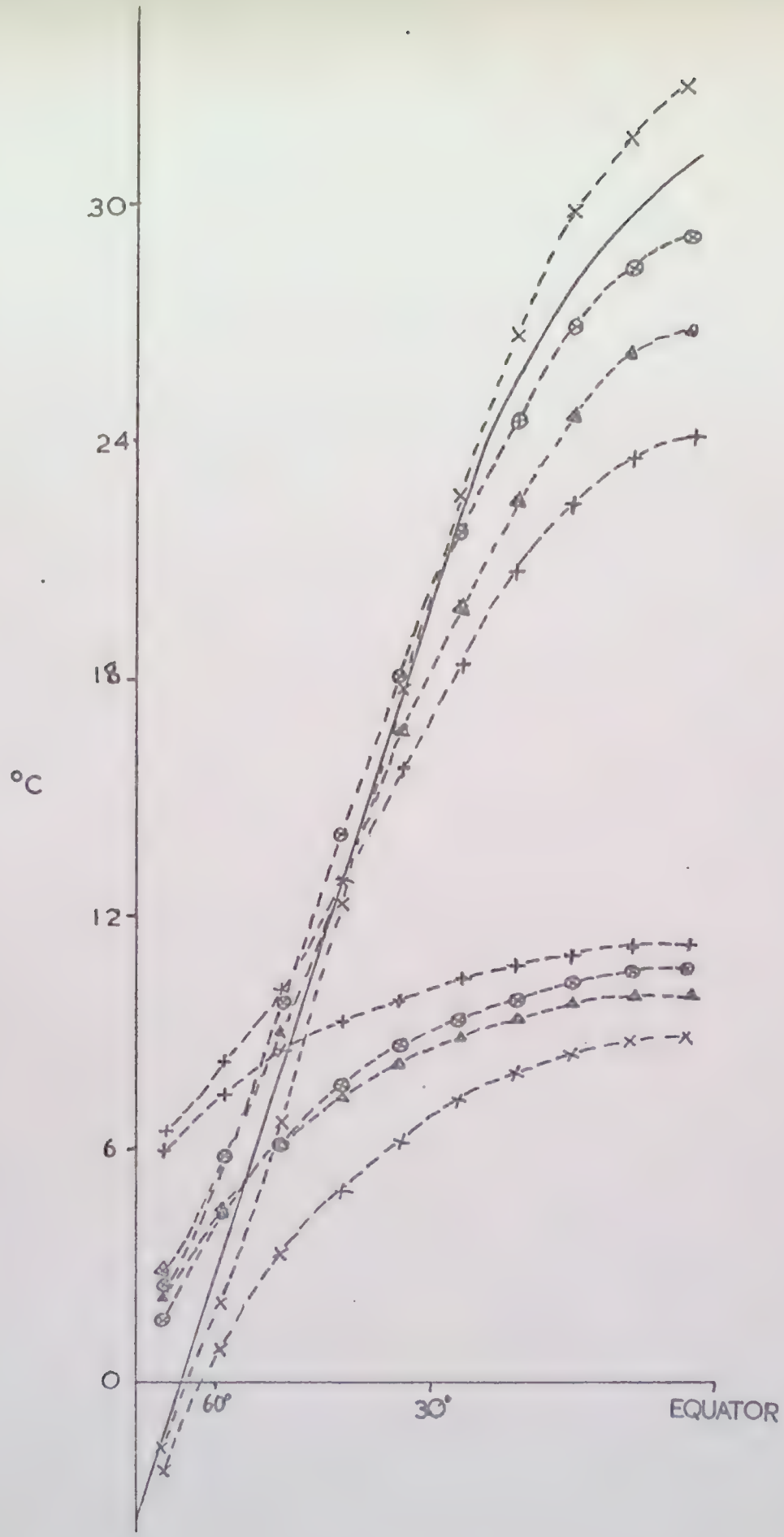

Figure 15. Meridional variation of SST and intermediate water temperature in the 3-level diffusive model with seasonal forcing and various values of the parameters λ_{md} (OML/intermediate water coupling constant), DD (intermediate water diffusivity) and D0 (combined diffusivity of OML and atmosphere).
+ λ_{md}=2, DD=3, D0=0.65; Δ λ_{md}=1, DD=1.5, D0=0.65; ⊗ λ_{md}=0.7, DD=1, D0=0.65;
× λ_{md}=0.7, DD=1, D0=0.35. (All units $Wm^{-2}K^{-1}$)
Solid curve shows solution for SST in 2-level model.

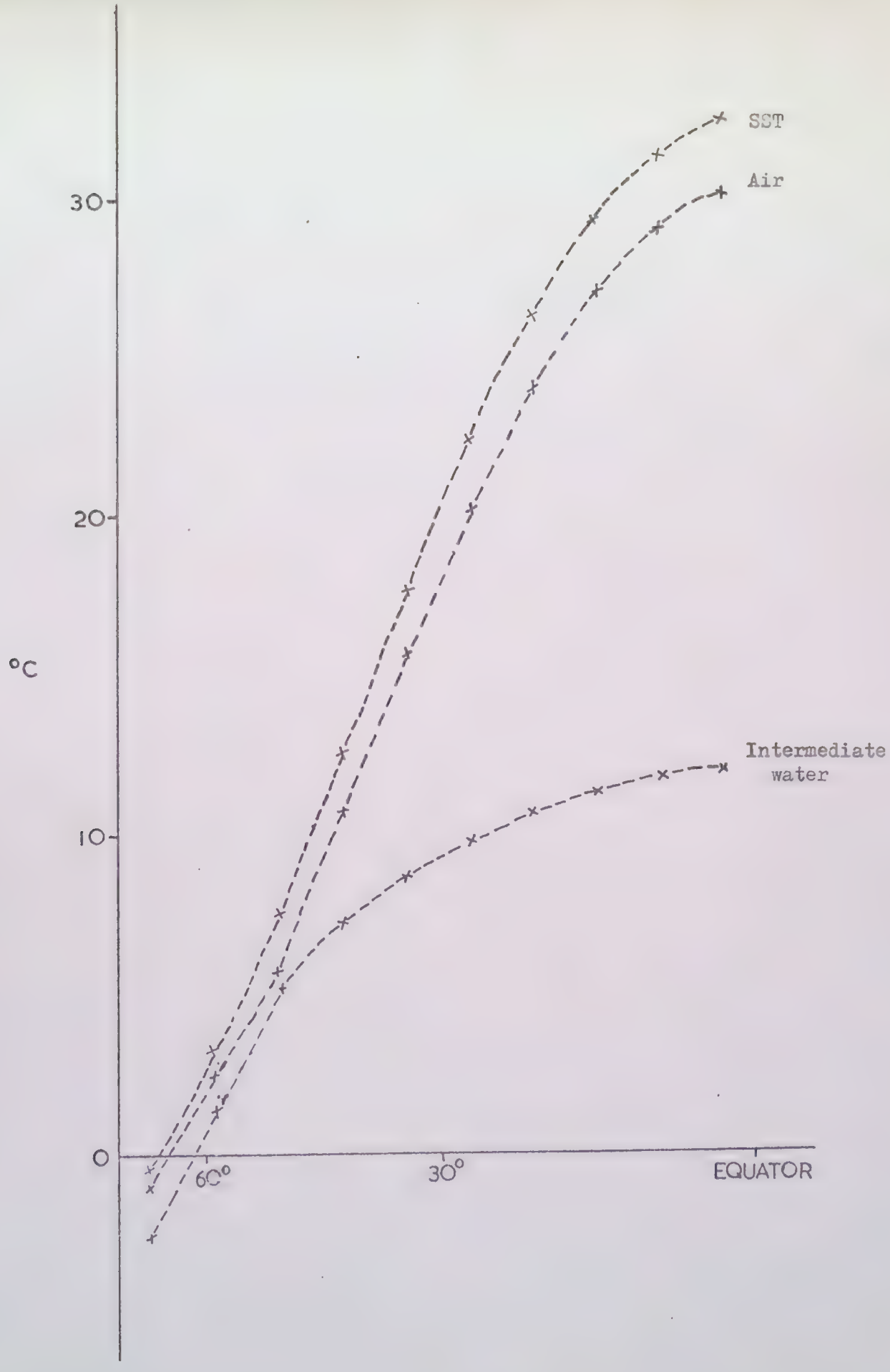

Figure 16. Values of air temperature, SST and intermediate water temperature in the 3-level diffusive model with seasonal forcing and D0=0.35, DD=1 and λ_{md}=1, taken after 1,100 years, when typical changes were 0.0001 K per century.

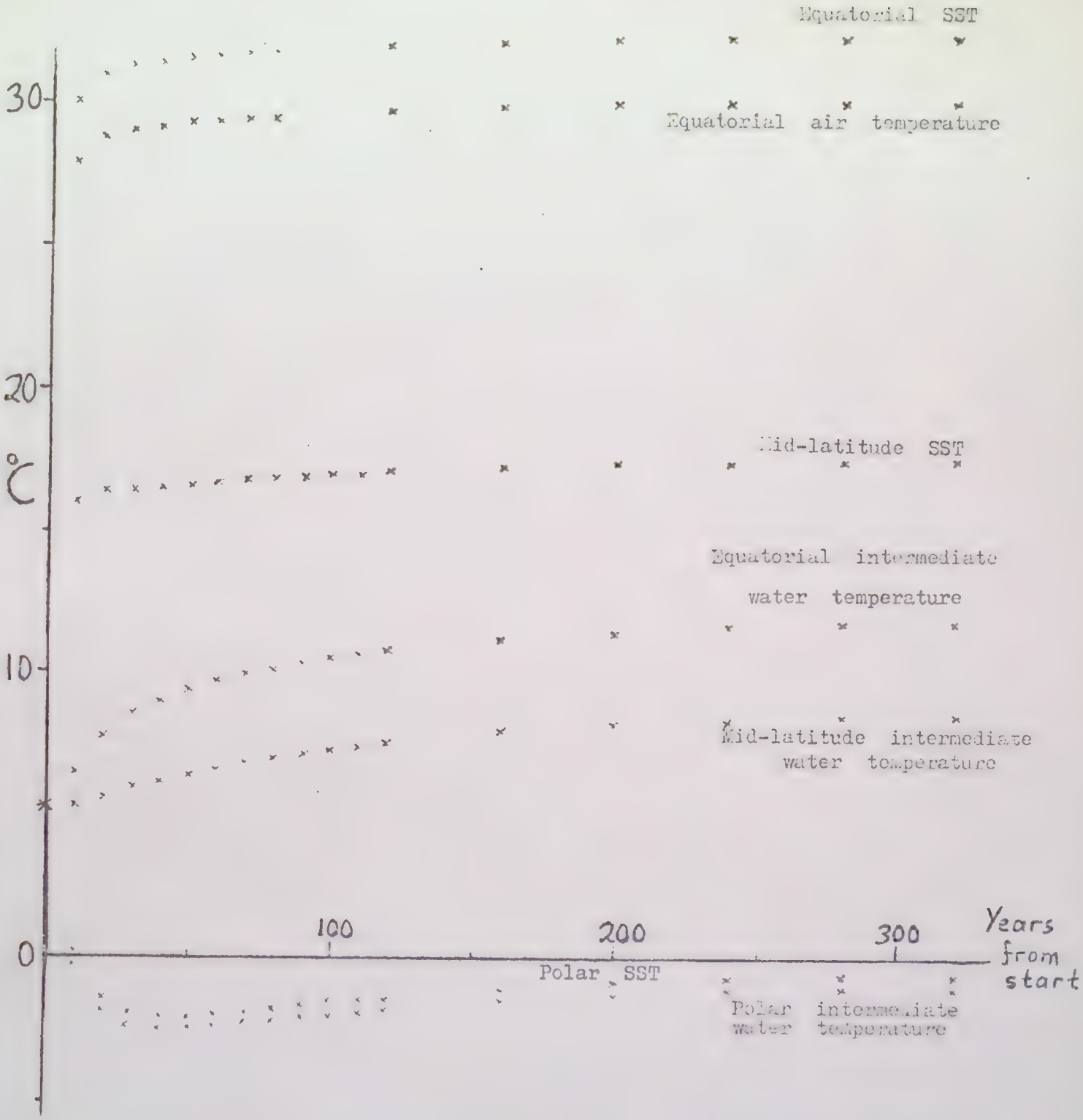

Figure 17. Evolution of SST and intermediate water temperature at three points at high (about 65o-90o), middle (about 30o-37o) and low (about 0o-6o) latitudes, together with air temperature at the former, for the 3-layer diffusive model with standard parameter values.

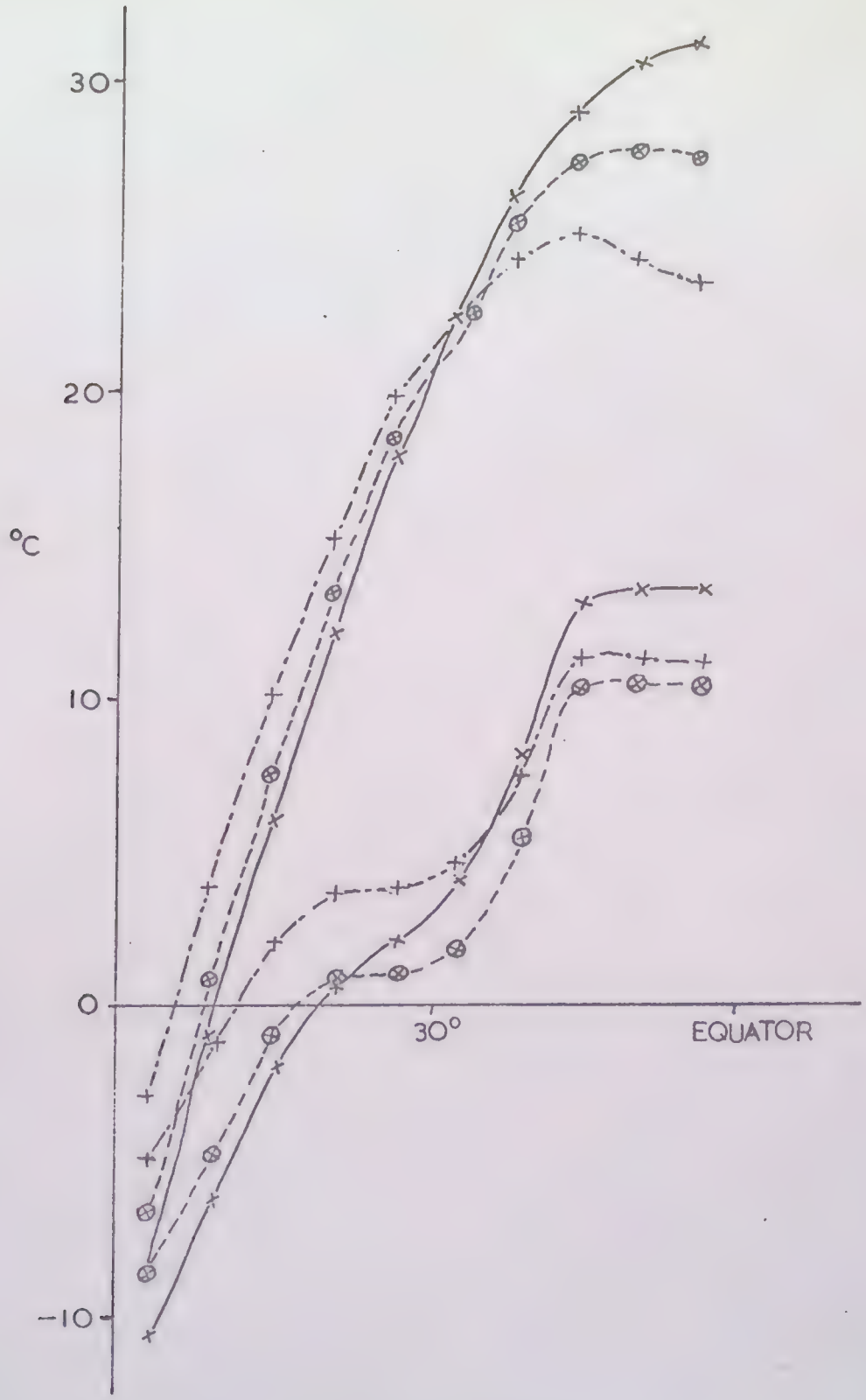

Figure 18. Zonal variation of SST and intermediate water temperature for the 3-level model with circulation for various strengths of circulation (⊗–⊗ standard strength, X——X 0.25 x this, +·–·–+ 2.5 x).

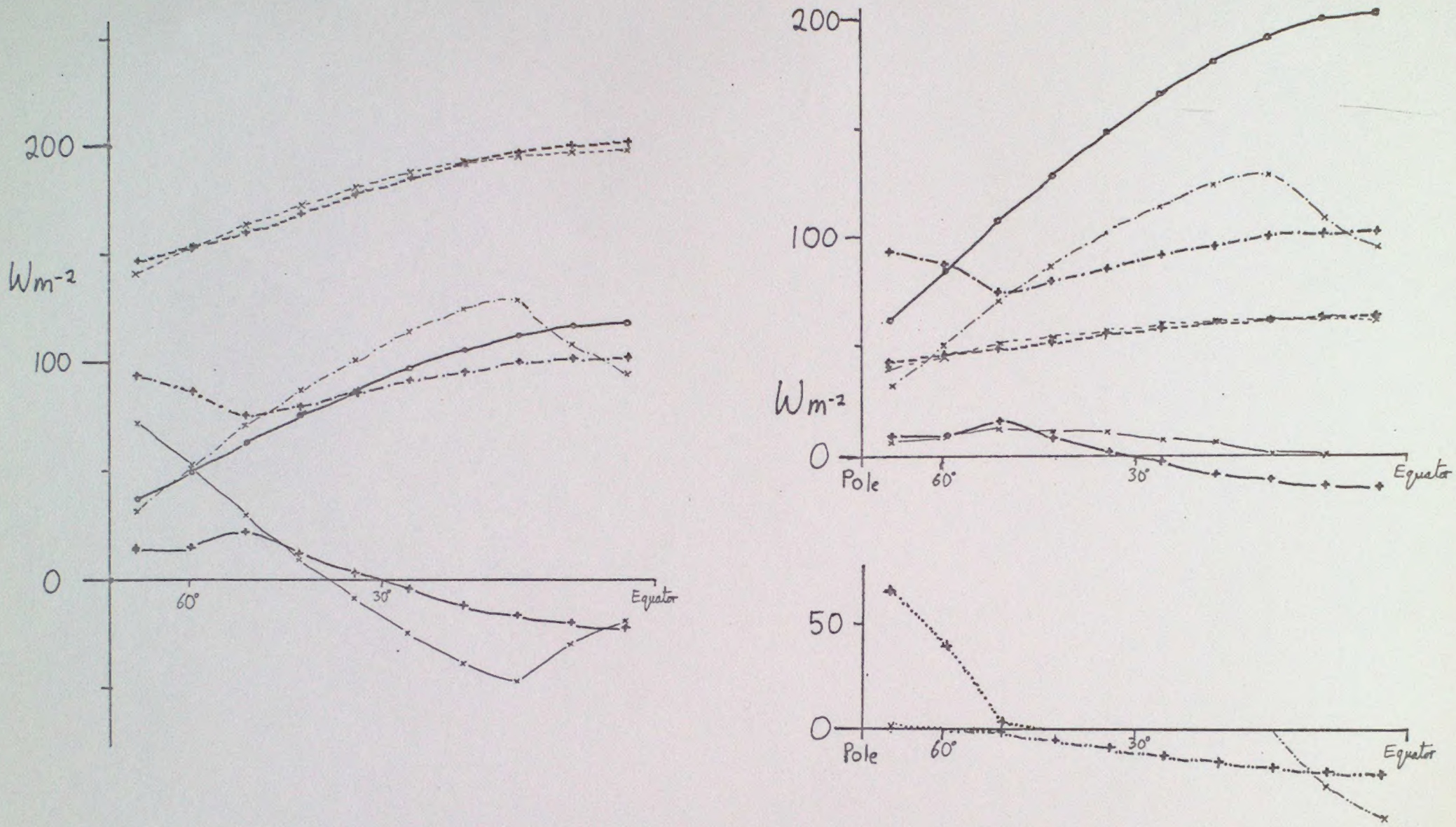

Fig 19. As for Fig 14, but for atmosphere (left) and OML (right). Symbols as for Fig 14.

Note that infrared cooling on both graphs, and sea-air exchange on the right hand graph are shown for convenience with the opposite of their physically correct signs.

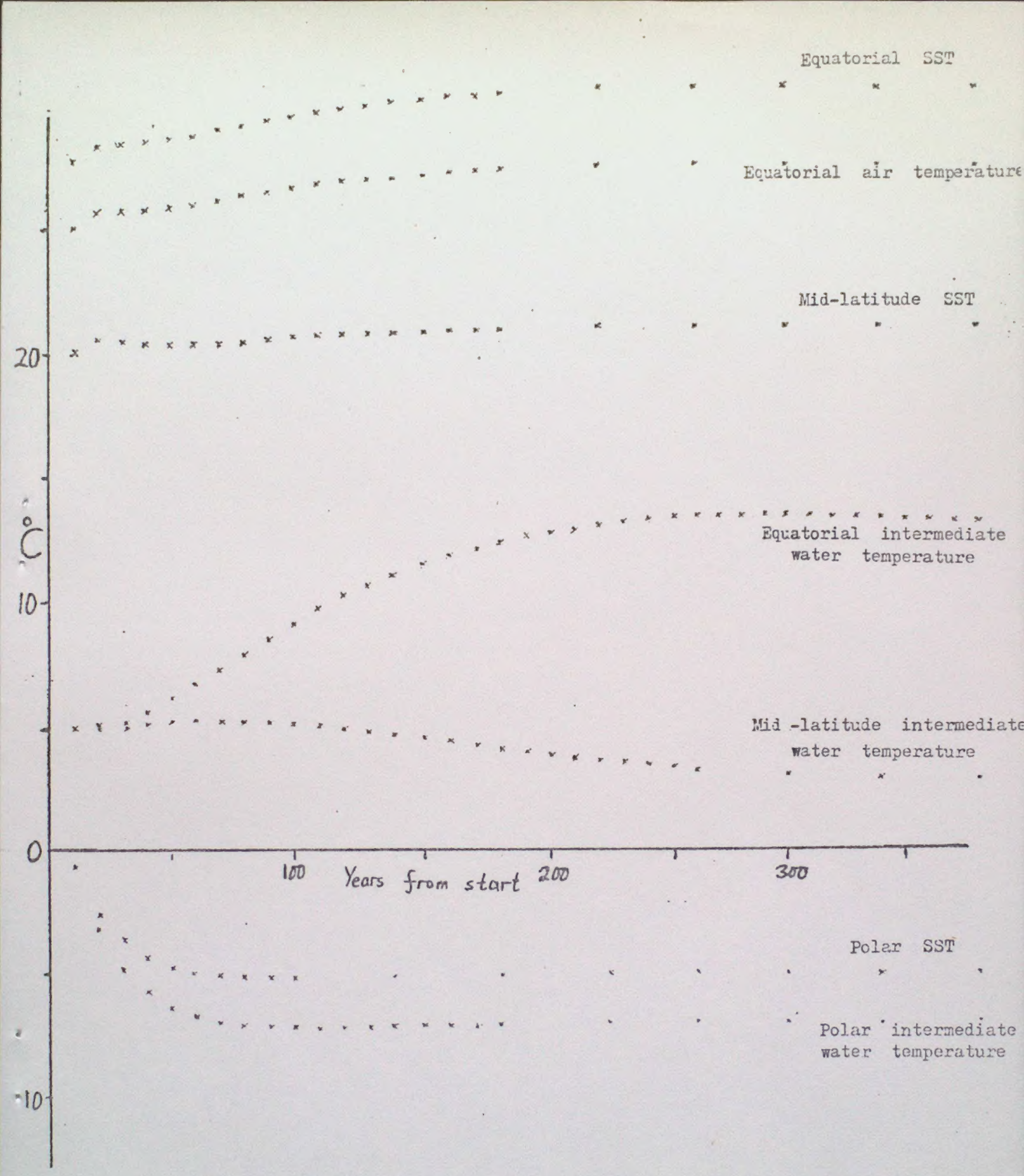

Figure 20. As for Figure 17, but for the circulating model.

DYNAMICAL CLIMATOLOGY TECHNICAL NOTES

DCTN 1	June 1984	A scheme for incorporating the thermodynamic sea-ice model into a coupled ocean-atmosphere model. C Gordon and M Bottomley
DCTN 2	Sept 1984	The parametrization of the upper ocean mixed layer in coupled ocean/atmosphere models. C Gordon and M Bottomley
DCTN 3	Sept 1984	The response of a simple thermodynamic sea-ice model to GCM forcing appropriate for perennial sea-ice. M Bottomley
DCTN 4	Sept 1984	Potential vorticity in the stratosphere derived using data from satellites. S A Clough, N S Grahame and A O'Neill
DCTN 5	Sept 1984	Simulation of the seasonal cycle of SST using globally specified climatological data to force simple ocean models. N S Grahame
DCTN 6	Sept 1984	Development of an improved GCM radiation scheme. A Slingo and R C Wilderspin
DCTN 7	Sept 1984	On the specification of surface fluxes in coupled atmosphere/ocean general circulation models. J F B Mitchell, C A Wilson and C M Price
DCTN 8	Sept 1984	A survey of possible repercussions of the Soviet river reversal programme on Arctic sea-ice. H Cattle
DCTN 9	Oct 1984	A study of a strong stratospheric warming during January 1982 T D A Fairlie
DCTN 10	Nov 1984	An intercomparison of the Met O 11 N45 model and the Met O 20 11-level model on the Cyber 205 J F Dyson
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DCTN 15	Jan 1985	Rossby wavetrains in the stratosphere forced by localised disturbances in the troposphere. C J Marks (Oxford), A O'Neill and V D Pope

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DCTN 16	Jan 1985	Observing system experiments using the Meteorological Office's 15-level model. R A Bromley
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